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GLACIAL HISTORY OF THE TROUT CREEK BASIN, SUMMERLAND, BRITISH COLUMBIA

by

DONALD RAYMOND KVILL

A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH

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THE UNIVERSITY OF ALBERTA

FACULTY OF GRADUATE STUDIES AND RESEARCH

The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research, for acceptance, a thesis entitled Glacial History of the Trout Creek Basin, Summerland, British Columbia submitted by Donald R. Kvill in partial fulfilment of the requirements for the degree of Master of Science

ABSTRACT

Ice stagnation in the upland areas commenced before, or contemporaneously with deglaciation of the topographic peaks. Because of this, moraines are absent and the till exposures that do occur are largely confined to the upland areas. Ice contact and ice-proximal features such as kames, kettled outwash and outwash trains occur within the valleys. These, together with numerous meltwater channels indicate that large masses of inactive ice downwasted within the valleys.

Deglaciation occurred as follows. First, the ice sheet that covered the study area stagnated in the upland areas and in the valleys that were oriented transversely to the direction of flow. As the ice thinned, stagnation became more widespread until eventually the lobe that occupied the Okanagan Valley became inactive. Meltwater produced by ice ablation was initially controlled by the slope of the ice surface but, as the ice surface downwasted, topography became increasingly important in channeling the flow. The exposure of lower meltwater outlets resulted in erosion of previously laid sediments by the streams to a lower base level which produced paired terraces along the sides of valleys such as Trout Creek. The final melting of ice from the south end of the Okanagan Valley produced a proglacial lake in which extensive silt deposits were laid down. These deposits were found to postdate the glaciofluvial material that was transported into the main valley via Eneas Creek.

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Thanks are also extended to the Mott family of Summerland who assisted me in various ways during the fieldwork phase of this study and provided me with an "oats-mobile" (or was it a Lincoln?) to roam the mountains on.

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Wives are often required to contribute a great deal toward their husband's ambitions and commonly receive little in return. Pat assisted me in the fieldwork phase of this study and has generously donated time from her own academic program to carry out many of the tasks that would otherwise conflict with the progress of the thesis.

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CHAPTER ONE

INTRODUCTION

The Okanagan Valley and adjacent areas of the Southern Interior of British Columbia clearly display the effects of glaciation. A reasonable account of the glacial history of the area can be obtained by the interpretation of the surficial deposits and the landforms within the area.

1.1 PURPOSE:

The objective of this study is to identify and describe the glacial landforms and surficial deposits within one small basin in the South Okanagan, and to construct an explanatory model of the deglaciation pattern of the area. Data were obtained from a two month detailed field study, an airphoto analysis and a literature review.

STUDY AREA: LOCATION AND DESCRIPTION

1.2 LOCATION:

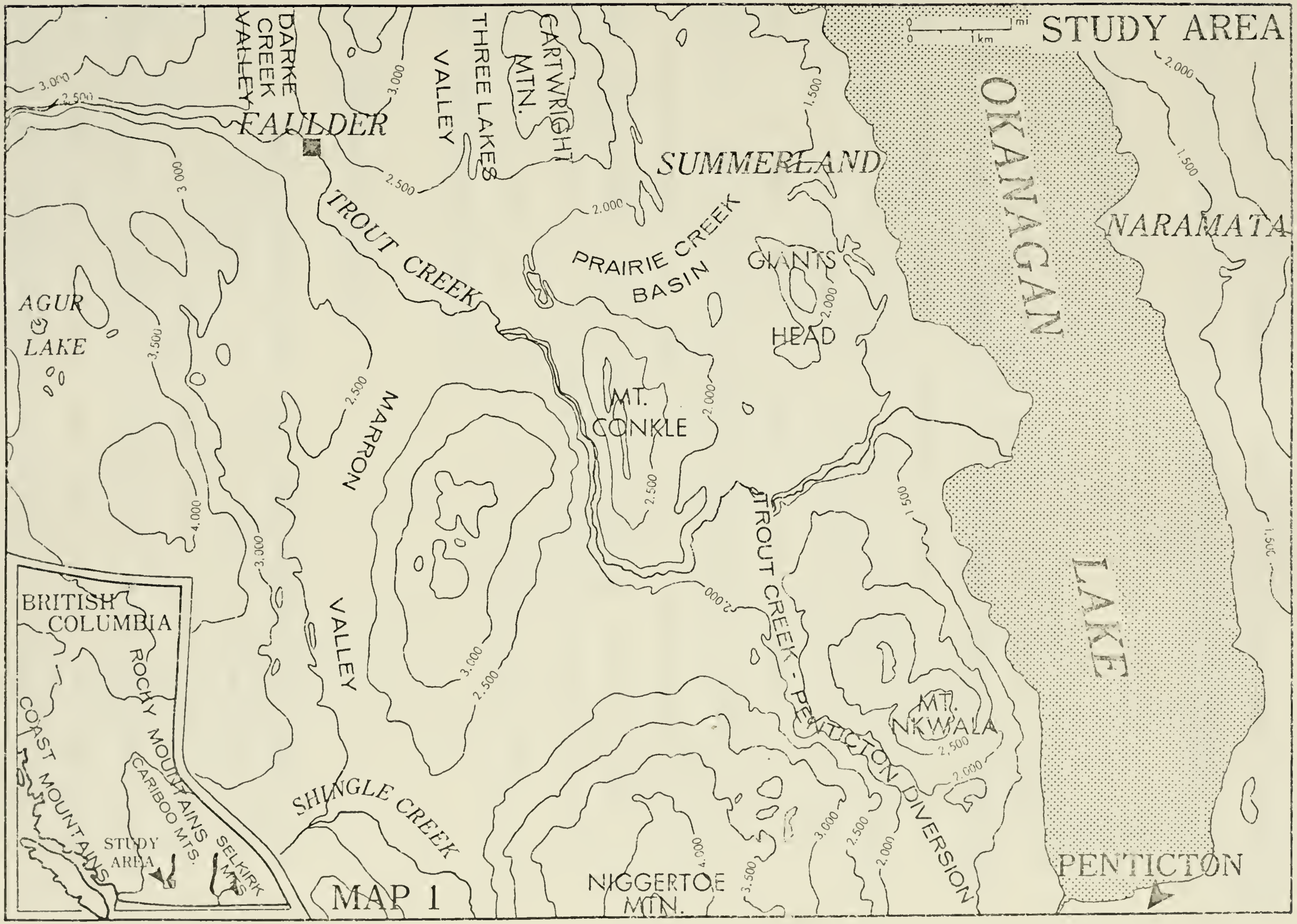
The Okanagan Valley lies approximately along the 119° 35' meridian of west longitude and is one of a series of north-south oriented valleys carved into the southern portion of the Interior Plateau of British Columbia. It extends from Shuswap Lake (lat. 51° N.) to the

Columbia River (lat. 48° N.). A second valley system, much smaller and often poorly defined, parallels the Okanagan from a point a few miles north of the study area to the International Border. This valley system is identified as the Marron Valley by Nasmith (1962).

Trout Creek Valley heads in the Thompson Plateau and extends eastward to form the Okanagan Valley in the vicinity of Summerland (Map 1). It transects the Marron Valley near the small settlement of Faulder. This confluence of valleys defines the centre of the study area. The Okanagan Valley and the Marron Valley are separated by glacially abraded mountains with a relative relief of approximately 450 m (1500 ft). The combination of peaks, intermediate valleys and lower main valleys allows the stratigraphic sequence of deposits to be traced from an elevation of 1375 m A.S.L. (4500 ft) to the present Okanagan Lake level of 342.3 m (1123 ft) A.S.L. The landforms of the main valley, the Marron Valley and the upland areas all correlate with those found within the Trout Creek Valley. Trout Creek transects all of the above mentioned features, thus allowing the stratigraphic sequence to be traced from the upland areas to the floor of Okanagan Valley.

1.3 BOUNDARIES:

The study area is defined by a large rectangle 306 km^2 (135 $mi.^2$) shown on Map 1. The northern side is an east-west line that includes the north end of Cartwright Mountain. The western boundary includes Agur Lake; the eastern side is located such that a minimum of three airphotos could be obtained for each of the flight lines along



STUDY AREA
0 mi
0 1 km
OKANAGAN
LAKE
NARAMATA
PENTICTON
SUMMERLAND
FAULDER
DARKE CREEK VALLEY
THREE LAKES VALLEY
CARTWRIGHT MTN.
TROUT CREEK
PRAIRIE CREEK
BASIN
GIANTS HEAD
MT. CONKLE
TROUT CREEK - PENTICTON DIVERSION
MT. NKWALA
MARRON VALLEY
SHINGLE CREEK
NIGGERTOE MTN.
AGUR LAKE
BRITISH COLUMBIA
ROCKY MOUNTAINS
CARIBOO MTS.
SELKIRK MTS
COAST MOUNTAINS
STUDY AREA
MAP 1
1,500
2,000
2,500
3,000
3,500
4,000

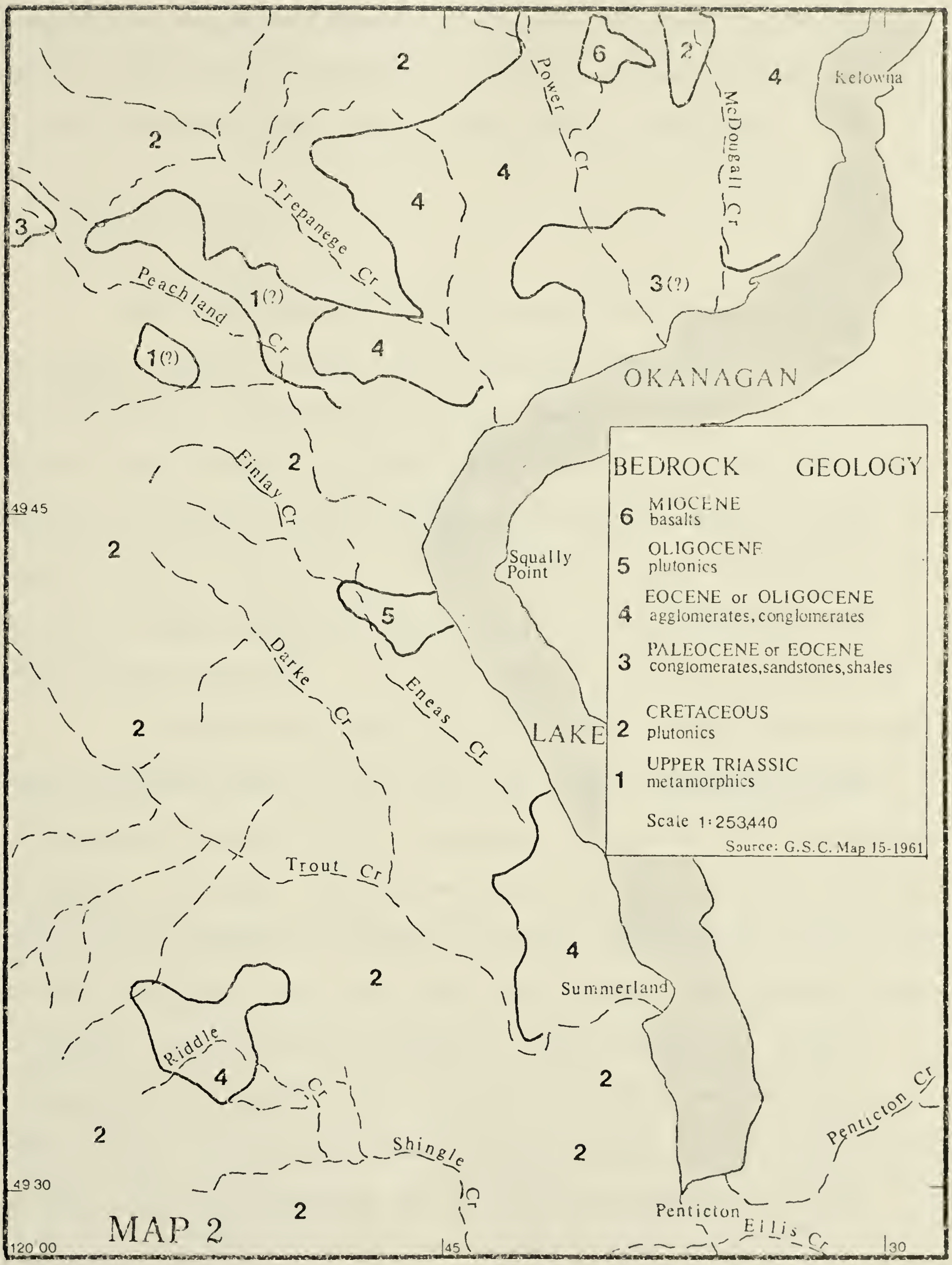
the east side of Lake Okanagan so that stereo coverage could be obtained. The southern margin is the south border of the 82E 12 Mapsheet.

1.4 BEDROCK GEOLOGY:

According to the most recent Geological Survey of Canada map (Map 15-1961), it appears that the Okanagan Valley is a structural trench overlying a system of subparallel faults (See Map 2). East of the Okanagan Valley, metamorphic rocks of the Monashee Group (probably Precambrian according to G.S.C. Map 15-1961) are prevalent whereas west of the valley younger (probably Cretaceous) plutonic rocks predominate. Two significant outcrops of Mid-Cenozoic volcanic rocks can be identified within the study area. One is located north of the lower end of Trout Creek and is exposed on Mt. Conkle and Giant's Head. The other is exposed near the headwaters of Riddle Creek in the southwest corner of the study area. North of the study area, near Kelowna, this unit outcrops more frequently.

1.5 PREGLACIAL DRAINAGE:

Kelley and Spilsbury (1949) and Fulton (1972) suggest that during the early Tertiary the Southern Interior Plateau region was a low, swampy area, the drainage from which was probably controlled by the same bedrock structure that controls the present drainage pattern. In the Middle Tertiary volcanic activity, tectonic uplift and extensive faulting and folding increased the local relief. This activity


Kelowna
Power Cr
McDougall Cr
Trepanege Cr
Peachland Cr
OKANAGAN
LAKE
Finlay Cr
Squally Point
Darke Cr
Eneas Cr
Trout Cr
Summerland
Riddle Cr
Shingle Cr
Penticton Cr
Penticton
Ellis Cr
49 45
49 30
120 00
45
30
BEDROCK GEOLOGY
6 MIOCENE basalts
5 OLIGOCENE plutonics
4 EOCENE or OLIGOCENE agglomerates, conglomerates
3 PALEOCENE or EOCENE conglomerates, sandstones, shales
2 CRETACEOUS plutonics
1 UPPER TRIASSIC metamorphics
Scale 1:253,440
Source: G.S.C. Map 15-1961

MAP 2

produced the topographic depression which was to become the Okanagan Valley. Soon after its formation, it became a drainage channel and by the end of the Tertiary a fluvial valley was well developed.

1.6 PRESENT CLIMATE:

The South Okanagan Valley has one of the mildest climates in Canada. Summers are warm with cool nights and winters are sufficiently mild that Okanagan Lake usually remains ice-free. A significant feature of the climate of the Okanagan Valley is that the temperature, aridity and number of frost-free days increases from north to south along the axis of the valley, and from higher to lower elevations across the valley (Okanagan Basin Agreement, Preliminary Report 38, Fig. D-1, D-2 and D-3).

A study of the thirty year (1941-1970) climatic record of the Summerland Experimental Station (B.C. Dept. Agric. Climatic Normals) indicates that the mean monthly temperatures are above 0°C for at least 10 months of the year. Overcast skies are prevalent during the months of November, December and January, but in the summer months, particularly in July and August, clear sunny skies are the norm. The overcast skies in winter, coupled with the proximity to Lake Okanagan, is largely responsible for the mild winter temperatures (Kelley and Spilsbury, 1949, p. 12). In July, the warmest month, the mean monthly temperature is 21°C. The extreme maximum for the thirty year record is a July temperature of 40°C. Precipitation at this station averages only 296.2 mm (11.66 in.) per year. The seasonal distribution of this

occurs in early summer, particularly in June, and mid-winter. The spring and fall months are characteristically arid with less than 23 mm. (0.9 in.) of precipitation in any month. When the precipitation data is compared with the 635 mm. (25 in.) of potential evapotranspiration that Warkentin (1967, p. 102) has calculated for the area, the extent of the aridity becomes evident.

1.7 VEGETATION:

Within the lower portions of the main valley moisture deficiency is sufficient to inhibit the development of forest vegetation. Instead, bunch-grasses, sage-brush and cactus occupy the areas that have not been irrigated. At higher elevations up the valley sides, and along the tributary valleys, montane forests are common. The dominant tree species are Ponderosa Pine (*Pinus ponderosa*) and Douglas Fir (*Pseudotsuga menziesii*). Usually the stand is open with bunch-grass and forbs occupying the open areas.

1.8 GLACIAL HISTORY OF THE SOUTHERN INTERIOR: LITERATURE REVIEW

The physiographic complexity of the Interior of British Columbia has resulted in an equally complex array of glacial landforms. Relatively little work has been carried out toward the development of a glacial chronology of the area despite the fact that postglacial aridity has preserved the glacial evidence and inhibited the development of a masking cover of trees. As a result the glacial history offered

in this review is generalized from a number of scattered and uncoordinated studies.

The number of glacial episodes that have occurred in the Western Cordillera is controversial. Early geologists (Dawson, 1887, 1890; Willis, 1898) working in the Southern Interior, postulated two advances based on two distinct till units separated by a unit of stratified sand and silt. Daly (1912), working in the Southern Interior along the 49th parallel, could not confirm this hypothesis. He suggested a single glacial advance but did not rule out the possibility of earlier glaciations, the evidence of which has been obliterated by the most recent advance. South of the study area, Flint (1935) recognized two glaciations in Eastern Washington based on drift deposits that had infilled earlier meltwater channels. Kerr (1934), working in north-central British Columbia, found evidence to suggest more than one advance prior to deglaciation but he was not certain that these represented separate glaciations rather than minor fluctuations of a single glacial period. Tipper (1971) found geomorphic evidence of more than one glaciation in the Central Interior and Fulton, (1975) has identified two drift layers in the Kamloops area and has provided radiocarbon dates (GSC-194; CGS-977 and others) which indicate that the earlier glaciation ended before 43,000 B.P. and the last advance began at approximately 19,100 B.P. Other workers have similarly found evidence for multiple glaciations in the mountain ranges that border the Interior Plateau (Johnson, 1926; Armstrong, _et al._, 1965; Hughes, _et al._, 1969) but to date no comprehensive regional model of deglaciation has been produced that can satisfactorily explain the observational data that has been amassed.

The Okanagan Valley is no exception. Nasmith (1962) has provided the only comprehensive glacial history of the area. Very little stratigraphic data are available. It is, however, almost certain that the Interior was extensively glaciated at least once prior to the last glaciation (Tipper, 1971), but the stratigraphic evidence of the earlier advance has largely been reworked.

CHAPTER TWO

RESEARCH TECHNIQUES:

2.1 FIELDWORK:

A surface reconnaissance was carried out on foot and horseback throughout the entire study area. The landforms were sketched onto the N.T.S. 82E/12 E & W maps (scale = 1:250,000). Because soil moisture retention is dependent upon texture, and texture is one of the criterion for classifying surficial deposits, this work was completed during the spring field session when the availability of moisture from snowmelt resulted in an observable vegetative response to soil drainage characteristics. This greatly assisted the identification of boundaries between surficial deposits.

2.2 MAPPING:

The structure, topography and bedrock geology of the study area have been mapped by the Geological Survey of Canada at a scale of 1:253,440 (Kettle River West sheet, Map 15-1961). The only mapping of the quaternary geology that has been done within the study area is the 1:126,720 map produced by Nasmith (1962). The mapping of the surficial deposits (Map 3) and the landforms (Map 4) was carried out primarily during the field reconnaissance. Critical areas were identified in

the field by the use of airphotos. Traverses were then plotted and a ground reconnaissance was carried out. As the surficial deposits and landforms were identified, the location was plotted on the airphotos. Where exposures were found, the structure, texture and imbrication of the material was noted. In areas that were ambiguous or critical, samples were taken and then returned to Edmonton for analysis. The deposits were classified on the basis of the depositional mechanisms and geomorphic agents. This information was interpreted from the texture and structure of the material and the topographic form of the deposits. The landforms were identified on the basis of composition, morphology and the relationship to other landforms. Frequently, the exact boundaries of the deposits and landforms could not be defined in the field. When this occurred, approximate boundaries were plotted on the airphoto in the field and the exact boundaries were defined later during the airphoto analysis.

2.3 SAMPLING

Field samples of tills, alluvial material and lacustrine sediments were collected for particle size analysis. Samples were taken in critical areas; the results of which are included in the interpretational work.

At each sampling site the soil horizons and organic layers were removed to expose the subsurface material. A visual inspection was carried out to determine the particle size range of the material so that a sample could be collected that would be at least 100 times the

weight of the largest particle. As sieves were not available for field sieving, the analysis of till units was limited. This is not a serious omission as the purpose in sampling the tills was to determine whether or not the finer sized particles have been removed by water.

Excellent stratigraphic sections were available in some of the terraces and kame deposits and these were logged and sampled where major regime fluctuations could be identified. Much of this work was purely descriptive, although imbrication and orientation studies were carried out on all exposures that included coarse gravels.

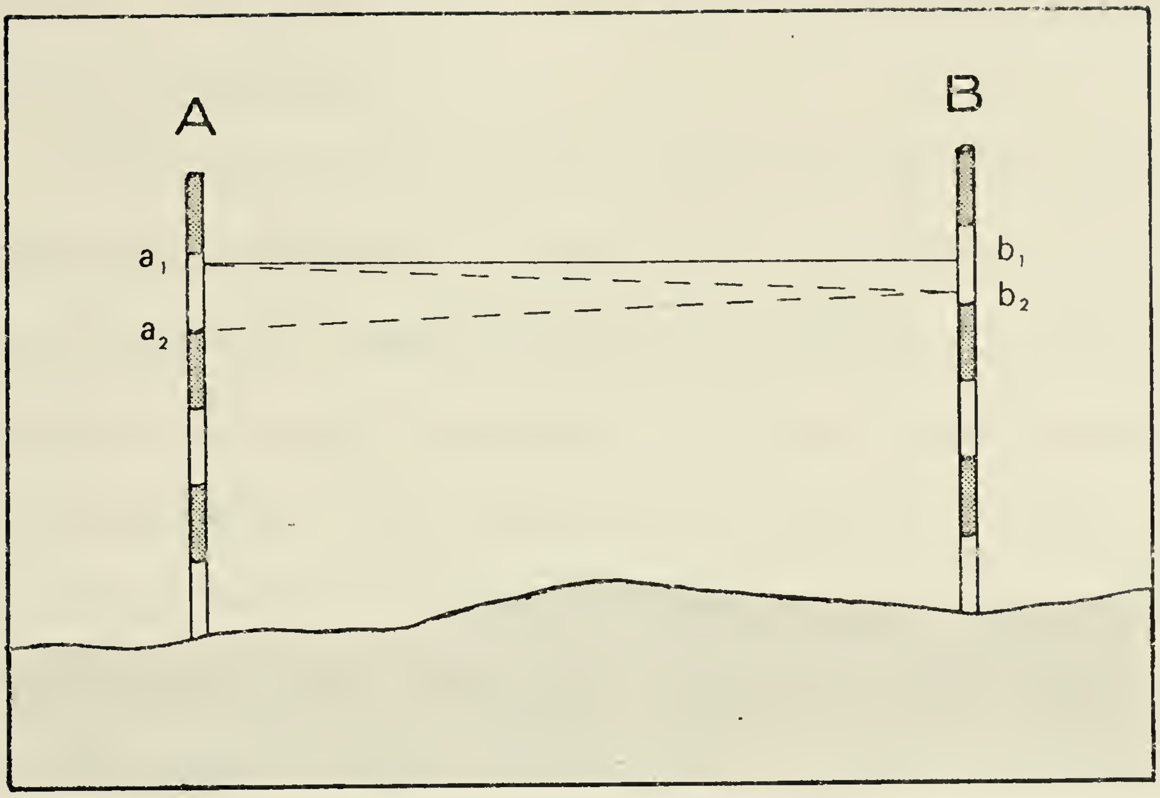
2.4 SURVEYING

A complete transect of the Trout Creek valley was made approximately 1.9 km. (1.2 mi.) southeast of Faulder. A Paulins Surveying Altimeter (Model M-5) was employed to determine spot elevations, which were located and mapped on the air photos.

The transect survey was carried out with the use of range poles and a Brunton compass (see appendix 1). A temporary benchmark was surveyed from the Geological Survey Benchmark 370-J. From this point, a traverse was surveyed from the valley floor to the uppermost definable terrace on either side of the valley. A maximum closure error of .45 m. was accepted over a horizontal distance of 1417 m.

In order to minimize instrument error, the Brunton Compass surveys were carried out in the following manner:

FIGURE 1



A foresight was taken from a datum point on pole A (point a) to establish a horizontal equivalent reference on pole B. Due to instrument error, the foresight will intercept pole B at some distance above or below point b_1. Suppose that the foresight intersects pole B at point b_2. From point b_2, a backsight is taken to pole A. The instrument error, if consistent, will result in a deflection such that the distance $a_1 - a_2$ is equal to twice the distance $b_1 - b_2$. If the range poles are measured from the ground surface, the location of point b_1 on pole B can be determined by using the formula

$b_1 = b_2 + \frac{a_1 - a_2}{2}$. The transect is continued by positioning pole C beyond pole B and repeating the sequence from point b_1. The horizontal distance was not measured directly along the transect. Instead, the survey points were plotted on the air photos from which the horizontal distances were calculated.

Several traverses across terrace sequences were carried out using the Paulins Altimeter for spot elevations and air photos for the horizontal distances. Stable daytime temperatures, low relative humidity and cloudless skies made it possible to obtain reliable results from the aneroid altimeter. All traverses were planned so that a base station reading could be obtained at least once per hour. A record of the elapsed time between reading was recorded so that slight changes in atmospheric pressure could be distributed. A variation in base station readings of three meters (10 ft.) was considered cause for rejection of the entire traverse. In most cases, the change in pressure was small enough to ignore.

2.5 LABORATORY TECHNIQUES:

2.51 PARTICLE SIZE ANALYSIS

Twenty-seven samples of 1000 to 2000 grams of material were returned to Edmonton for laboratory analysis. All of the samples were dry sieved at one-half ϕ (phi) intervals down to 3.5 ϕ (0.088 mm.). The material finer than 3.5 ϕ was dispersed with a 5% solution of sodium hexametaphosphate $Na(PO_3)_6$ and wet sieved to remove the 4.0 ϕ

fraction. Phi values for the material finer than 4.0 ϕ were obtained by pipette analysis. It was considered necessary to adopt a technique by which the material finer than 4.0 ϕ could be delivered to the sedimentation apparatus without first oven drying the sample to obtain an accurate weight of the material being delivered to the sedimentation test. The reluctance to oven dry the sample was based on the difficulty of disaggregating the material which would still retain the sodium hexametaphosphate. Mechanical disaggregation would possibly induce a bias in the size distribution of the sample.

The method adopted was to split and weigh a subsample of the material finer than 3.5 ϕ. This subsample was then dispersed and passed through the 4.0 ϕ sieve and the material greater than 4.0 ϕ was dried and weighed. The percentage of the material that had passed through the sieve could then be calculated, assuming no loss of material on the sieve. It was reasoned that if 10 g of material finer than 4.0 ϕ were required for the pipette analysis, the split weight of material finer than 3.5 ϕ that would deliver the required 10 g could be calculated from the proportions recorded in the subsample test. Due to the difficulty of splitting very small fractions of material, any weight of material that would deliver between 8 and 10 grams of sample to the pipette analysis was accepted. An accurate weight of material delivered could then be calculated for the pipette analysis. The assumption of no weight loss in the wet sieving was demonstrated to be unwarranted when the 8 inch diameter sieves were used. In order to control the losses, a 3 inch diameter sieve was

designed and constructed. This sieve could be installed in the top of a 600 ml beaker with a press fit. In addition, the total weight of the sieve was 73 g, so it could be weighed on a balance scale before and after each sieving.

The adoption of this technique allowed for an uninterrupted transition through the wet sieve into the sedimentation apparatus and losses were minimized by fewer transfers of material.

A standard pipette analysis technique was adopted with 10 ml samples being removed from 10 cm beneath the surface at time intervals equivalent to 5.0 ϕ, 5.5 ϕ, 6.0 ϕ, 7.0 ϕ, 8.0 ϕ, 8.5 ϕ, and 9.0 ϕ. This extended the particle size data to silt/clay boundary.

2.52 AIRPHOTO INTERPRETATION

The most significant criterion for the identification and interpretation of surficial deposits from airphotos is the form or morphology of the unit because topographic expression, soil characteristics such as texture and structure, and spatial relationships are diagnostic of the processes involved during deposition. For this reasons, extensive use was made of remote sensing techniques in the development of the landform map and the geomorphic interpretation.

One hundred and ten black-and-white panchromatic images with a scale of 1:15,840 (1 in = 1320 ft) were obtained from the university collection. In addition, ERTS images were purchased at a scale of 1:1,000,000 from the National Air Photo Library. The airborne imagery was studied and mapped at the Alberta Centre for Remote Sensing

where an Interpretoskop (trade name), a micro-densitometer and a zoom transferscope are available.

(i) Interpretoskop

This instrument is essentially a binocular mirror sterioscope that has a zoom magnification capability, controls to correct for image distortion due to variation in the flight line, and a 'floating dot' for measurement of vertical heights from stereo pairs. The eye-base distance can be varied to alter the vertical exaggeration of surface topography and the illumination of either photo can be varied to compensate for differences in tone between the individual photos.

On the basis of field notes and the analysis of airphoto data, a landform map was produced. The surficial features were mapped directly onto the photographs.

(ii) Zoom transferscope:

The transferscope is an optical instrument which superimposes the image of an airphoto onto a base map. Controls are available to enlarge, shrink or stretch the airphoto image in any direction such that it can be made to fit known reference points on the base map. The zoom feature enables the operator to superimpose images of various scale combinations. In addition, it enables adjustments to be made to compensate for radial distortion or changes in scale that result from changes in relief. The illumination on the photo or the map can be

altered to compensate for changes in tone and contrast between the imagery and the base map. This instrument was used to transfer the landform data from the airphotos to a base map with a scale of 1:125,000.

(iii) Micro-densitometer:

The analysis of tonal variations is one of the major photographic keys available to the interpreter. The micro-densitometer enhances the tonal variations with the aid of a video camera and screen. The photograph is scanned by the video cameraand the grey scale is broken down into approximately 2.5 times as many shades of grey as the human eye is capable of discriminating. False colors are added to any or each of these tones as it is being projected on the screen.

It must be remembered that tonal variations can be influenced by changes in sun angle, changes in the albedo of the surface, film processing techniques and lense aberrations. These factors limit the utility of the diagnostic value of the technique. The interpreter must be sufficiently familiar with the target area to be able to select only the data that are pertinent to the study. Evaluation studies of this technique have shown that tonal change and photographic texture can be valuable for the interpretation of landform features, the comparison of landforms in different stages of dissection, and lithologic comparisons (Estes & Senger, 1974). Ray and Fischer (1960) demonstrated that the textural variations that exist between glacial moraines of different ages and between morainic and alluvial deposits

can be identified by tonal variations on aerial photographs.

The densitometric study of the photo coverage of the study area furnished evidence to support boundary locations that were questionable using only the Interpretoskop. It was hoped that the technique would be of value in defining the upper levels of the pro-glacial lake that formed in the south end of the Okanagan Valley (Glacial Lake Penticton - Nasmith, 1962) but the agricultural and cultural patterns have masked the surface texture patterns.

CHAPTER THREE

DESCRIPTION OF SURFICIAL DEPOSITS AND LANDFORMS

The surficial deposits were identified and mapped on the basis of texture and sedimentary structure (see Map 3). The landform units were interpreted on the basis of morphology, composition and the location relative to associated landforms (see Map 4).

3.1 SURFICIAL DEPOSITS

3.11 TILL

Till is material deposited directly by ice. It is non-stratified, not obviously sorted and contains fractured and angular clasts (Flint, 1971, pp. 152-154). Till deposits within the study area exist only in the form of ground moraine. The particle size data on the 3 till samples indicates a generally sandy textured matrix with 75% to 90% of the sample coarser than 4 ϕ. However, some of the till examined contained boulders up to several meters in diameter. The lithology of the pebbles, cobbles and boulders was almost invariably granite. Scratch marks and crescent-shaped gouges exist on many of the larger boulders.

3.12 LACUSTRINE DEPOSITS

Sediments that have accumulated in nonflowing or very slowly moving water bodies are defined as lacustrine deposits.

The composition of the lacustrine deposits within the study area consists of fresh feldspathic rock flour, usually cream-white to pale buff in color (Flint, 1935, p. 110) and is stratified in parallel laminae that are commonly 10 cm to 15 cm thick. Massive laminae as thick as 200 cm were measured. The texture of these deposits was typically fine sand, silt and clay (see Appendix 1). Rhythmites found in glacial lake sediments are not always varves, but Embleton and King (1968, p. 438) suggest that most of them can be interpreted as such. Varve data presently being studied indicates that these lacustrine deposits were deposited in less than 200 years, (M.J. Kent, pers. com.). The thickness of the rhythmites, therefore, indicates the rate of sediment yield from the ablating glacier and by counting the number of lamina the lifespan of the lake can be determined. The normal size for glacial lake rhythmites is 1 cm to 5 cm thick according to Embleton and King (1968, p. 437). It is therefore evident that the rate of sedimentation within the Okanagan Valley was much greater than normal. The lacustrine deposits that bracket the southern end of the valley display evidence of sagging and slumping. Flint (1935) concluded that no evidence of compressional folds or thrusts is evident, but normal faults are common, particularly adjacent to the cliff side of the deposits.

3.13 ALLUVIAL DEPOSITS

These are deposits laid down by postglacial streams. Within the study area they take the form of alluvial fans and recent floodplains. They can be distinguished from similar glacial deposits by the fact that they are graded to the present level of Lake Okanagan.

The alluvial deposits consist of moderate and well sorted and stratified sands and gravels containing some silt. The clasts are well-rounded and range in size from cobbles with a 20 cm (7.8 in) "a" axis to silt-sized material. The floodplain of Trout Creek above Mt. Conkle has a typically sandy texture, whereas the alluvial fans within the main valley are composed of coarse gravels and sand overlain in places by a thin mantle of silt.

3.14 GLACIOFLUVIAL DEPOSITS

Sorted and stratified material deposited by streams issuing from a melting glacier are identified as glaciofluvial deposits. They are differentiated from recent fluvial deposits by their topographic position and the base-level to which they are graded. Five of the landform units are composed of glaciofluvial deposits. These are kames, kame terraces, outwash plains, fans, (late glacial and alluvial) and terraces.

Kames Steep-sided glaciofluvial deposits that were deposited in direct contact with an ice surface are identified as kames. Meltwater was the depositional medium, hence the deposits are sorted and stratified.

The ice proximal environment is responsible for the significant variations in texture and sorting that are observable in vertical exposures.

The kame that transects the upper end of the Prairie Creek Valley has these characteristics. An exposure at the south end of the ridge complex displays moderate to well sorted, horizontally stratified beds of coarse sand with occasional pebble-sized inclusions. Along the west side of the complex is an exposure which is composed of well-rounded granite boulders and cobbles interbedded with sands and gravels (see plate 2). To the north, the sediments grade into a rapidly alternating sequence of coarse gravels and sand (see plate 3). Along the eastern side, (ice contact side) of the kame, the upper 2 m - 4 m of the deposit consists of a nonstratified facie composed of a mixture of sand, silt and clay (see plate 4). This unit is only found at the top of the highest ridge and is probably diamicton that flowed directly from the ice. Beneath this facie are laminae of sands and gravels. The degree of sorting and stratification varies from well sorted and stratified in some laminae to poorly sorted and stratified in others. Some of the beds along the eastern side of this exposure shows evidence of collapse as the beds now are dipping to the east while the paleo-current direction generally indicates a flow toward the southwest.

<u>Kame Terraces</u> Kame terraces consist of fluvial deposits that have been laid down along the floor of streams that have occupied

the channel formed between the side of a stagnant or nearly stagnant ice lobe and the valley wall. The composition and structure reflects this environment in that coarse sand and gravel are the most common textural classes, but all textural sizes from well-rounded cobbles up to 25 cm (10 in) in diameter to fine sand can be found. Kettles and collapse features can be observed along the abandoned ice-contact surfaces. The material is usually well rounded and derived from granitic parent material. Cross-bedding and ripple structures are common within the deposits. Kames and kame terraces are indicative of slowly melting stagnant ice, (Embleton and King, p. 386; Flint, 1971, p. 212).

Outwash Trains Outwash trains consist of sorted and stratified material that has been carried away from a glacier by meltwater streams. Compositionally, this material differs from those of kame terraces in that no deformation has occurred within the sediments and the texture range and structural variations are less dramatic which is indicative of the fact that deposition occurred at some distance from the ice front. Well-rounded gravels and cobbles predominate in these deposits, which are almost entirely of granitic material. Cross-bedded ripple structures and cut-and-fill bedding (as per Flint, 1971, p. 186) are frequently observable and the coarser textured material is imbricated furnishing a means by which the paleocurrent direction can be determined.

Within the Okanagan Valley outwash material has infilled the glacial trough to an incredible thickness. Seismic studies carried out by the Geological Survey of Canada indicate that the bedrock floor of the

northern end of the valley, north of Vernon, lies 183 m (600 ft) below sea level (G.S.C. Paper 72-8). Borehole studies indicate that the 535 m (1350 ft) of fill is composed primarily of sands and gravels (Fulton, 1972). A second borehole that has not yet been interpreted was drilled in the south end of the valley near Okanagan Falls for the British Columbia Water Investigation Branch. Here, 232.5 m (763 ft) of sands and gravels were logged before bedrock was reached. An approximate log of the borehole data is as follows, (Fulton, pers. com.).

Depth	Material
0' - 188' (0 - 57.3m)	Coarse bouldery gravel, possibly cemented
188' - 220' (57.3m - 67m)	Coarse to fine gravel
220' - 420' (67m - 128m)	Sand & gravel with silt interbeds
420' - 490' (128m - 149.3m)	Coarse gravel, minor sand with silty interbeds
490' - 510' (149.3m - 155.4m)	Silt?
510' - 540' (155.4m - 164.6m)	Sand & gravel
590' - 710' (164.6m - 216.4m)	Sand & gravel interbedded with silt
710' - 750' (216.4m - 228.6m)	Gravel
750' - 763' (228.6m - 232.6m)	Sand, minor gravel
763' - 818' (232.6m - 249.3m)	Till, possibly weathered
818' - 822' (249.3m - 250.5m)	? Gravel
822' - 848' (250.5m - 258.5m)	Tertiary conglomerate

(Source: Photocopy of correspondence between B.C. Water Investigations Branch and Dr. Fulton, G.S.C.)

Terraces Terraces are remnants of outwash or fan material that has been eroded to a lower base level. Terrace deposits are compositionally similar to outwash and fan deposits in that they consist of stratified sediments that freq·ently show evidence of cross-bedding or ripple structures which are useful for the interpretation of the depositional environment. Good exposures were found in the well-defined terraces that bracket the Trout Creek valley between Faulder and Mt. Conkle. The lower facies consist of laminated or cross-

bedded sands, but the top 1 to 2 m consists of coarse, well-rounded boulders and cobbles between 15 cm and 80 cm in diameter. This has been interpreted as a lag deposit and it is partly on this basis that they have been identified as terraces, rather than kame terraces.

Fans Fans are found where streams emerge from upland areas and enter into a larger valley. At the junction of two valleys, a decrease in gradient or a widening of the stream channel results in the deposition of sediment. Compositionally, fans consist of moderately sorted and poorly stratified alluvium or glaciofluvial material. Cross-bedding is common and frequently inclusions of fine textured material can be found in an otherwise coarse gravel or cobble-sized deposit. Frequently, the texture of the surface material fines with increased distance from the apex of the fan.

3.2 LANDFORM UNITS

3.21 ICE SCOURED BEDROCK

Ice abrasion of the Southern Interior has produced landforms that bear similarities to areas that have been eroded by ice sheets or ice caps as well as alpine areas where differential glacial erosion was prevalent. The upland areas of the Southern Interior have been rounded and subdued by ice erosion whereas the major valleys, particularly the north-south oriented ones, have been widened, straightened and deepened into the classic U-shaped cross-profile. The magnitude of erosion that took place during the last glacial advance is difficult

to determine. The rounded and abraded nature of the upland areas and the over-deepening of the Okanagan Valley are evidence that extensive erosion has taken place but as yet it is not known how much of this is attributable to the last major advance.

Most of the ice-abraded surfaces are masked by glaciofluvial deposits or till and therefore are not shown on the landform map (see map 4). The bedrock outcrops that do exist within the study area are easily identifiable in the field and on airphotos. Generally, they exist along the margins of the valleys and along the northern end of the upland areas. Crag-and-tail topography (Flint, 1971, p. 103) is common. Mount Conkle is an excellent example with exposed bedrock above 761 m (2500 ft) A.S.L. along the north end and a bevelled mantle of till masking the flanks and southern (lee) side. The abrasion pattern on Mt. Conkle indicates abrasion from an ice mass moving due south. Striations can be identified on the north side of this feature from an elevation of approximately 610 m (2000 ft) A.S.L. to the summit (845 m, 2774 ft). Measurement of the orientation of the striations at eight different locations on Giant's Head, plus four sites at an elevation of 550 m (1800 ft) in West Summerland supports the hypothesis that ice movement in the Summerland area was approximately north to south.

The incidence of ice-scoured bedrock is greater along the east side of the Okanagan Valley. Whalebacks (Flint, 1971, p. 97) are the most common morphologic expression and these are eroded in a manner that indicates a direction of ice movement from a north-easterly

direction (018^{o} - 020^{o} T). This indicates a flow of ice that is approximately 35^{o} east of the general orientation of the main valley which signifies that the ice was not deflected around Squally Point but rather flowed directly south from Kelowna to the east of Okanagan Mountain. Northeast of Pentiction a large whaleback with local relief of more than 60 m (200 ft) exists near the valley floor. The orientation of this feature is approximately 011^{o}T which is again approximately 24^{o} east of the axis of the valley.

Small-scale stoss-and-lee features are common on the north and east facing slopes of the upland areas. The orientation of some of these features was measured at an elevation of approximately 975 m (3200 ft) A.S.L. along the west side of the upper Deschamps Creek Valley (see Map 4). It was found that they indicate an ice movement from 005^{o}T. Similar landforms were measured on the north end of the mountain between Three Lakes Valley and Darke Creek (Meadow) Valley at an elevation of approximately 915 m (3000 ft) A.S.L. and an orientation of 005^{o}T was again obtained.

An ice-polished cliff exists between 855 m (2800 ft) and 975 m (3200 ft) at the northwest end of Meadow Valley, just to the north of the study area. This indicates heavy abrasion by ice which was deflected into Meadow Valley over a divide which is approximately 732 m (2400 ft) A.S.L. It appears that when the ice in the main valley was above the level of the divide, a substantial amount of ice flowed into this valley which is oriented such that it provides an exit to the south for some of the ice that was deflected toward the west side of the valley by Squally Point. The remainder of the ice was

forced to navigate the sharp bend in the valley at this point.

The absence of classic erosional landforms is significant. Horn peaks, arêtes and cirques are not present within the study area. A very careful search of the upland areas failed to identify so much as a nivation hollow. This implies that after the ice sheet thinned sufficiently to expose the upland areas, the climate was sufficiently mild to prevent perennial snow patches at these elevations. It is possible that such landforms were formed during earlier interglacials or interstadials but the erosion of the last advance appears to have removed all evidence of their existence.

3.22 GROUND MORAINE:

In respect to morainic landforms, the ones that cannot be found are again the most significant. No terminal, lateral or recessional moraines were found, indicating that no major still-stands or minor readvances interrupted the deglaciation of the area.

Within the Okanagan Valley ground moraine occurs only at elevations above 460 m (1509 ft) A.S.L. while at lower elevations glaciofluvial and lacustrine deposits are found. In the Marron Valley the 730 m (2395 ft) contour line usually marks the contact between till and the water-deposited sediments. Three Lakes Valley is a noteable exception. In this valley till is found throughout the valley except for minor lacustrine and fluvial deposits along the axis of the valley.

It is possible that till underlies the stratified deposits that occupy the lower elevations in the study area but at present

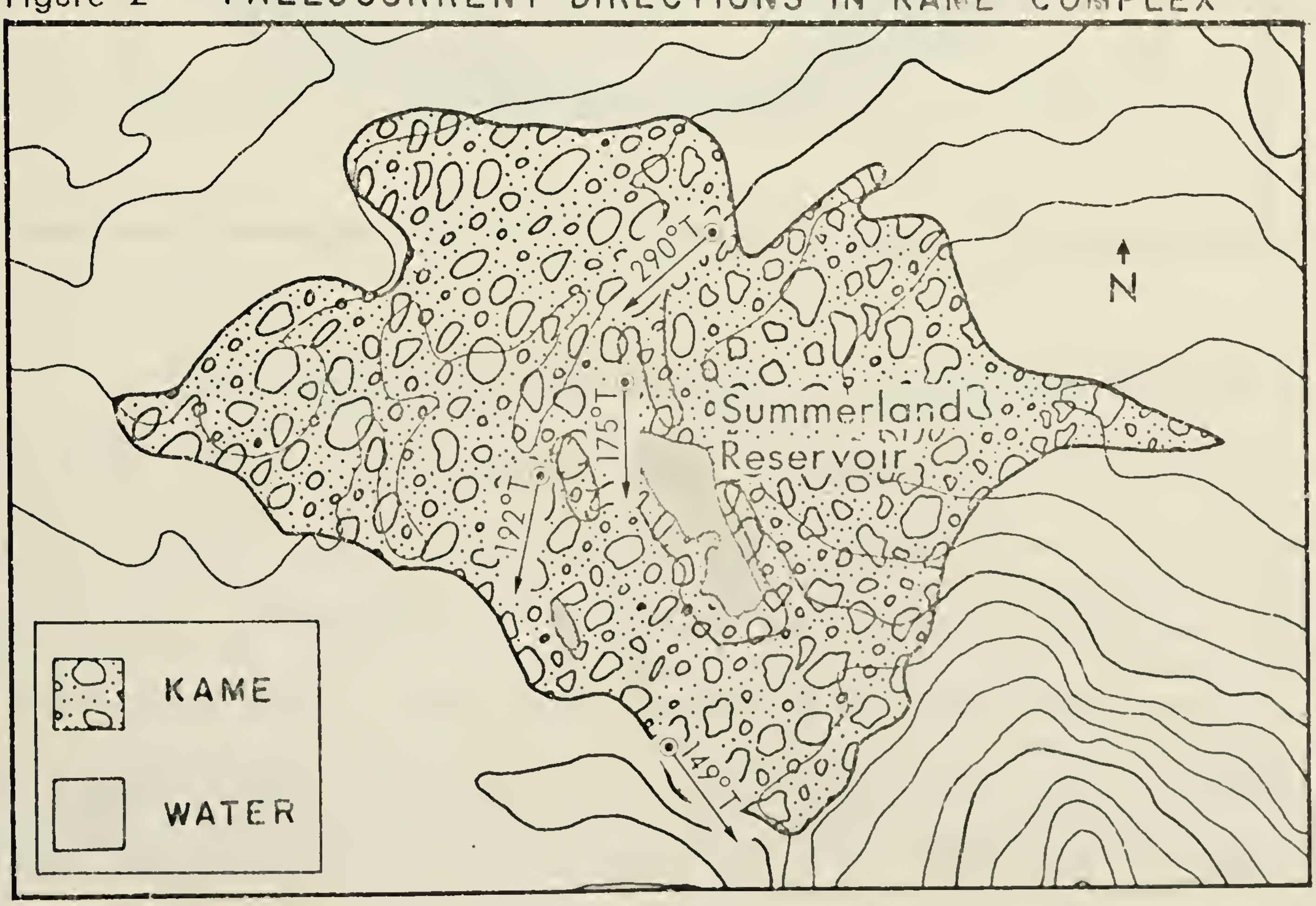
this has only been confirmed in two locations. Approximately 1.7 km (1 mile) southeast of Faulder till is exposed in a recent cutbank of Trout Creek. Stratigraphically this unit underlies all of the glaciofluvial deposits within the Trout Creek Valley. R.J. Fulton (Geol. Surv. Can.) has identified a till unit at the base of the unconsolidated deposits that occupy the bottom of the Okanagaon Valley near Okanagan Falls (south of the study area). The partly completed interpretation of a well log indicates the presence of a till unit 17m (55 ft) thick which is overlain by 232.5 m (763 ft) of glaciofluvial material (see borehold log p. 25).

3.23 KAMES:

Only one kame was mapped. It transects the valley between Mt. Conkle and Mt. Cartright at the upper end of Prairie Creek. Kames are composed of stratified drift and include at least one steep side which slopes approximately at the angle of repose for sand and gravel. The kame in question (see plate 1) consists of four distinct sub-parallel ridges plus minor incomplete ridge fragments. There is a slight curvature to the ridge pattern such that Prairie Creek is on the concave side and Trout Creek is on the convex side. Both sides of the ridges are approximately at the angle of repose (about 42°). The upper surface of these ridges is irregular with a slight north-to south gradient. An elevation of 646 m (2120 ft) A.S.L. was measured as the highest point. Immediately to the east (concave side) of the feature is the Prairie Creek Valley which has been mapped as thin

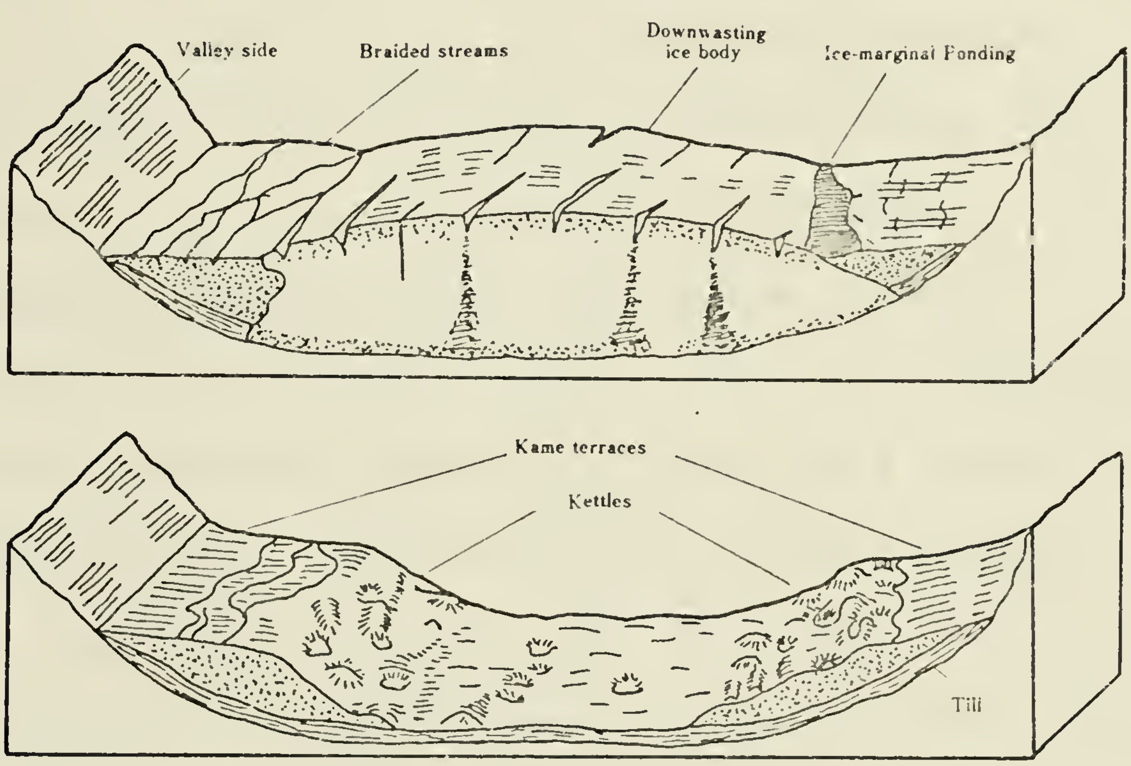
glaciolacustrine material over bedrock. It appears that an ice lobe extended into this valley from the Okanagan Valley during the active phase of glaciation and it stagnated in the Prairie Creek Valley with the distal end extending to the western side of the kame ridge complex. It is postulated that the release of pressure as a result of reduction of ice thickness by melting, plus differential melting along the shear planes of the glacier caused fluvial sediments to accumulate in the shear plane crevasses, thus creating the transverse ridges. Final melt-out of the ice allowed the sediments to collapse to the angle of repose which accounts for the discordant ridge heights and irregular valley morphology. Several paleocurrent measurements were obtained from sedimentary structures such as ripple marks and from imbrication studies. The results of these investigations are as follows:

Figure 2 PALEOCURRENT DIRECTIONS IN KAME COMPLEX



3.24 KAME TERRACES:

Figure 3: Formation of Kame Terraces (after Flint, 1971)



Kame terraces are found in pairs or coelescent groups along the valley walls in analogous positions to lateral moraines in classic alpine landform associations. The Okanagan Valley, Trout Creek Valley, Marron Valley and Meadow Valley all have kame terraces along some portion of the valley walls. Typically, they appear as flat-topped terraces with a steep riser facing the axis of the valley. The upper surface, in some instances, has a slight slope toward the valley axis. Where

the tread is almost horizontal, the upper surface frequently shows evidence of having been a braided streambed. Kettled topography is almost invariably associated with kame terraces.

The kame terraces in the northwest corner of the study area are highest in elevation. They bracket the southern end of Darke Creek (Meadow) Valley and are found along the north side of the Upper Trout Creek Valley at an elevation of approximately 790 m (2600 ft) A.S.L. Further down the Trout Creek Valley, kame terraces occur along the valley sides as far as the junction with the Marron Valley. The upper surface of this terrace assemblage is generally at 701 m (2300 ft) and a measured elevation of 721 m (2367 ft) was obtained for a point in the intersection of the two valleys. Within the Okanagan Valley, kame terraces are found at elevations between 458 m (1500 ft) and 488 m (1600 ft) A.S.L. Ground moraine or ice-scored bedrock is invariably found along the valley walls above these features and outwash trains, terraces or lacustrine deposits occupy the lower elevations.

3.25 OUTWASH TRAINS:

Outwash trains are linear deposits of glaciofluvial sediments. Where confined by valley walls, these deposits could be defined as valley trains but in the upland regions outwash deposits exist that appear to have been deposited from streams that were largely supraglacial. The upper surfaces of the outwash trains are commonly flat with only minor irregularities in the micro-relief which are suggestive of an abandoned braided stream pattern (see plate 5). Commonly kettle holes

exist in the upper surface of the outwash material.

Outwash trains are proglacial deposits therefore the existence of these landforms indicates that at the time of deposition the area in question was ice-free while up valley,or at lower elevations,ice may have still been present.

Outwash trains are found at different elevations. The highest, and therefore the earliest in terms of the deglaciation sequence, is an extensive but poorly defined deposit along the west side of the study area. It is found at an elevation of approximately 1158 m (3800 ft) A.S.L. and is deeply kettled. Agur Lake and the small unnamed lakes adjacent to it now occupy kettles within this outwash train (see map 4). The Okanagan Study Group Report 15 states that Agur Lake is 7 m deep and at least 20 m of outwash material can be observed above the water level which suggests that the deposit is at least 27 m (88.6 ft) thick. This indicates that a large amount of meltwater was being produced at the time that the upland areas were being exposed.

At a slightly lower elevation (approximately 762 m (2500 ft) an outwash train complex can be identified along the west side of the Marron Valley (see map 4). The northern (up valley) end of this complex is heavily kettled and the degree of kettling decreases rapidly in a downvalley direction (see plate 5). The uppermost level of this complex is about 1.2 km (3/4 mi) south of Faulder. Downvalley from the kettles, the upper surface becomes very flat and is interrupted only by some shallow, smooth-sided channels one to two meters (3.3 ft - 6.6 ft) deep that are oriented in a down-valley direction.

A north-south transect was surveyed along 505 m (1660 ft) of the upper surface of this feature and it was found that the slope was toward the south at 1° (see plate 7).

An outwash train also lies along the floor of the Marron Valley. It begins about midway down the valley and the upper elevation at the northern end is 670 m (2300 ft) A.S.L. Shingle Creek has eroded away the deposits from the confluence of the Shingle Creek and the Marron Valley but it is evident that the outwash train originally transected the Shingle Creek Valley and continued south along the Marron Valley. Dissected remnants of the old outwash still exist at the confluence of the two valleys.

3.26 LACUSTRINE PLAINS:

Lacustrine plains are flat, or nearly flat, deposits of silt and clay. Where these deposits are found along valley walls, extensive post-glacial gullying has occurred producing steep-walled ravines that dissect the otherwise featureless benches. The term lacustrine plain has been applied to all lacustrine features so as to avoid complicating the mapping criteria with distinctions that are relatively unimportant in terms of the interpretation. Lacustrine plains are indicative of ice-marginal or proglacial lakes that formed during deglaciation.

The lacustrine plains within the study area can be explained using one of three possible models. The landform assemblages and the topographic relationships between the landforms are used to determine

which model is the most plausible for any specific deposit. The three models of glacial lake development are:

1. Proglacial lakes which form in front of a retreating ice lobe where the local topography obstructs the removal of meltwater. In some cases the bedrock slopes toward the retreating ice front and in other instances dams of glacial drift have been transported into the valley and obstructed the drainage. It is thought that both of these mechanisms were involved in the glacial lake that occupied the southern end of the Okanagan Valley.
2. Ice marginal basins are frequently created by a lobe of ice within a larger valley which obstructs the drainage of smaller tributary valleys. The lacustrine deposits which occupy the sides of valley walls such as along the Upper Trout Creek Valley are of this sort.
3. Kettle-hole lakes occupy depressions that contained blocks of stagnant ice during deglaciation. The resulting topographic hollows may have temporarily filled with meltwater. Some of these kettle holes are presently water-filled but others, such as the depression in the Marron Valley just south of Trout Creek, have drained and no longer are capable of retaining water, given the present climatic conditions.

Within the Marron Valley a lacustrine plain exists a short distance south of the Trout Creek-Marron Valley junction (see map 4). It lies between a deeply-kettled kame terrace to the north and an

outwash train to the south. The lacustrine plain is lower than either of the two adjacent features and there is no drainage channel out of the basin except for a small gully into the adjacent kettles down-valley. The location of this feature in a natural constriction of the Marron Valley, plus the above mentioned characteristics suggests that a block of ice stagnated in this location. This lacustrine plain was therefore formed according to the third model above.

A lacustrine plain occupies the bottom of Prairie Valley and the small valley immediately west of Giant's Head, between the elevations of 548 m (1800 ft) and 503 m (1650 ft) A.S.L. Bedrock outcroppings and a slightly irregular surface morphology indicates that these deposits are very thin. The kame ridge complex at the upper end of Prairie Valley and the till deposits between the higher bedrock exposures at the southern end of the valley west of Giant's Head suggests that a lobe of ice extended into this area and melted in situ. As it melted back proglacial ponding occurred as described in model one above.

The most extensive lacustrine plains are found bracketing the southern end of the Okanagan Valley. They appear as steep-faced, flat-topped benches 40 to 45 m (131 - 150 ft) high adjacent to Lake Okanagan (see plates 9 and 10). The cliff faces of these benches display clearly the sedimentary structures referred to above (3.12). The upper surface of the benches slopes towards the axis of the valley at 3°-5° for the most part. Flint (1935) however, cites slopes as steep as 8°-10° in local areas but his description of these steeply sloping

areas and the observations made during this study suggests that he is including outwash fan deposits from the tributary valleys as part of the lacustrine terraces. In addition, Flint states that the maximum width of the lacustrine benches is found at Summerland where it is about 2 miles (3.2 km) wide. Field observations indicate that this is not so. At Summerland it was observed that the lacustrine deposits were laid down against the foreset beds of a delta that extends from Eneas Creek. The contact between the two deposits is less than 0.8 km (1/2 mi) from the present lakeshore (further discussed in 3.29).

The interpretation of these lacustrine plains is difficult. Flint (1935, p. 114) and Nasmith (1962, p. 41) suggest that a long tongue of stagnant ice with a relatively flat surface occupied the axis of the Okanagan Valley during deglaciation and the lacustrine deposits were laid down partly on ice and partly against the valley wall in a manner analogous to kame terraces. The problem with this explanation is that it requires the existence of a tongue of ice at least 25 km (16 mi) long but only 3.2 km (2 mi) wide (the present width of Lake Okanagan) and some 153 m (500 ft) thick (present lake bottom to the top of the lacustrine benches). It seems very implausible that such a needle of ice would last for a significant length of time and remain at the bottom of the valley with extensive lakes lying along both sides of it. Perhaps the ice was sufficiently laden with debris that it would not float. The question then is how could it remain relatively static for between 100 and 200 years (see 3.12).
years (see 3.12).

The most critical evidence against the existence of this finger-of-ice model is the existence of a second complete terrace that occupies the bottom of the valley approximately 15.2 m (50 ft) beneath the surface of Lake Okanagan (Okanagan Study Group, Rept. 18, p. 10). This terrace has been traced down both sides of the valley from Squally Point to the south end of the lake and within Skaha Lake. Prograding beach deposits and longshore drift deposits have been identified in association with this feature (ibid., p. 10).

A more probable explanation for the existence of the lacustrine benches would be that because of the sharp deflection of the valley around Squally Point, the ice in the southern end of the lake stagnated and melted out much earlier than the ice north of Squally Point. Heavy ablation plus an ice or drift dam at some point south of the study area, possibly between Okanagan Falls and McIntyre Bluff as Nasmith (1962, p. 41) suggests, created a proglacial lake from Squally Point south. Rapid infilling of the valley by sediments resulted in the eventual breach of the restriction to the south and rapid drainage of the ponded water incised a canyon into the lacustrine sediments. After the high discharge period was over, a lake formed within the valley at a level approximately 15.2 m (50 ft) below the present lake surface. Wave action against the fine textured sediments created the lower terrace and the associated beach deposits. For some reason, probably aggradation of the postglacial fans that coelesce in the valley floor beneath Penticton, the lake level was raised to its present height. Wave action again trimmed the base of the lacustrine

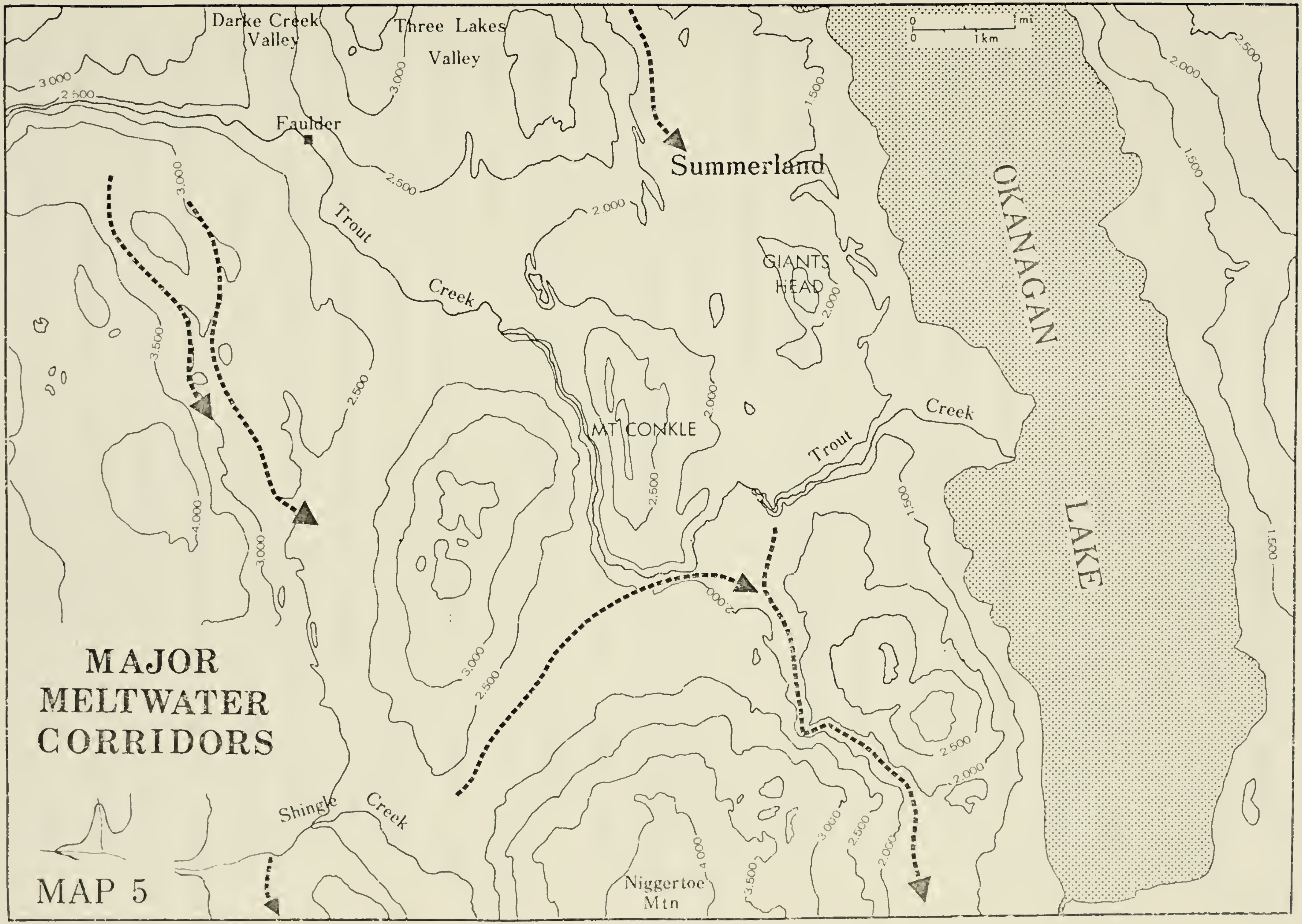
sediments thus producing the near-vertical cliffs observed along both sides of the lake today.

Additional evidence, such as core samples from the subaquatic terraces or a more detailed investigation of the structure and composition of the lacustrine benches, may one day furnish the critical evidence required to confirm this hypothesis. Until then the issue will remain somewhat speculative.

3.27 MELTWATER CHANNELS:

Meltwater channels are recognizable stream-cut channels that result from meltwater erosion. Within the study area they generally contour valley walls or transect upland areas and show no evidence of having been occupied by postglacial streams (see plates 11, 12). In some instances the ice margin provided one wall of the stream channel and the valley wall the other. In other areas proglacial drainage has been diverted through existing topographic depressions. The elevation of meltwater channels is an approximate indication of the elevation of the ice surface at the time of occupancy. The gradient of ice-marginal meltwater channels is an indication of the slope of the ice surface according to Embleton and King (1967, p. 275).

The number of meltwater channels within the study area is so great that no attempt was made to map each of them. The upper surfaces, east and west flanks, and the north-facing exposures of the upland areas above an elevation of about 760 m (2500 ft) A.S.L. are commonly grooved by meltwater channels. The orientation of some of the larger



MAJOR MELTWATER CORRIDORS

MAP 5

meltwater channels was measured and it was found that the general direction of flow was toward the south but with a deflection away from the main valley. This seems to indicate that the surface of the ice lobe within the Okanagan Valley was higher than the adjacent areas. Meltwater was therefore routed away from the main valley via minor valley systems such as the Marron. From this we could interpret that the main valley was occupied by active ice after the upland areas were becoming exposed through an essentially stagnant ice mass. Alternatively, it might only reflect a more rapid ablation in the areas adjacent to the exposed uplands and the entire ice mass may have been relatively inactive.

Tracing the meltwater pattern within the study area provides evidence for the sequence of deglaciation. Therefore, the individual meltwater channels will be described from the highest to the lowest elevations providing a rough chronologic sequence. During the earliest stages of deglaciation the drainage pattern cannot be traced due to the fact that the entire area was covered by ice and the flow would have been supraglacial or englacial. As the uplands became exposed, erosion and deposition by meltwater began. The outwash deposits around Agur Lake (see 3.26) are depositional landforms from these earliest meltwater streams. As the ice surface lowered, topographic barriers became more important in channeling the meltwater flow and ice-marginal channels were formed. The first of these was a system of interconnected and sometimes discontinuous ice-marginal channels incised into the upland area southwest of Faulder. The

elevation of these channels varies from 1220 m (4000 ft) down to 915 m (3000 ft) A.S.L. During the early stages of the occupancy of this system a meltwater channel was carved across the interfluvial upland between Riddle Creek and Shingle Creek. As downwasting progressed, successively lower channels were incised until the large canyon at elevation 915 m (3000 ft) was occupied. This canyon presently exists as a steep-sided gorge carved into bedrock of the western wall of the Marron Valley to a depth of more than 90 m (300 ft).

A small meltwater channel exists along the axis of Three Lakes Valley. Nasmith (1962, fig. 3) suggests that this valley was a meltwater corridor of major significance during the earlier stages of deglaciation. Meltwater, he suggests, flowed south through this valley and upon reaching the Trout Creek Valley it deflected westward, because of the ice blockage at the lower end of Trout Creek, and resumed a southward flow along the Marron Valley. The fieldwork for this study produced little evidence to support this hypothesis. The valley is mantled by a thin layer of till and only a small meltwater channel exists along the axis of the valley.

The meltwater channel mapped along the northwest side of Niggertoe Mountain requires some qualification. It was mapped as a meltwater channel due to the fact that the most recent geomorphic event related to the deglaciation appears to have been the flow of water from the southwest toward Trout Creek. The unconsolidated deposits along this valley are relatively thin, especially toward the northern end where abraded bedrock bluffs interrupt the glacial deposits. Widespread, but thin deposits of lacustrine material are

found along the valley bottom on either side of the meltwater channel. It would appear that a tongue of ice extending up the lower Trout Creek Valley obstructed drainage from this area until a very late stage in the deglaciation. Proglacial ponding occurred against the front of the ice lobe and lacustrine sediments were deposited, these can be observed along the sides of the present valley. As downwasting of the ice obstruction continued, an exit for the meltwater was exposed into the Trout Creek-Penticton Diversion. A short inter-connecting meltwater channel indicates that such a flow did exist but there is little evidence of the flow that Nasmith suggests which would empty into the present Trout Creek Valley. Nasmith (1962) suggests that this valley acted as a watershed (as opposed to a meltwater) drainage channel and this interpretation is very plausible.

A significant meltwater channel already mentioned is the Trout Creek-Penticton Diversion mapped by Nasmith (1962). This channel connects the Trout Creek system with the Okanagan Valley from a point south of Mt. Conkle to the west side of the main valley near Penticton (see map 4). It is relatively low in elevation (550 m, 1800 ft A.S.L.) and is adjacent to the Okanagan Valley which indicates that it formed at a late stage in the sequence of deglaciation. Like the meltwater channels along the western side of the Marron Valley, this channel consists of a deep gorge carved into bedrock, the walls of which were cut so steeply that the valley floor can no longer be observed in the deeper sections due to the accumulation of postglacial colluvium.

3.28 TERRACES:

Some of the most impressive features within the study area are the excellently preserved terraces found along the sides of the present-day streams, abandoned meltwater channels and alluvial fan streams (see plates 13 and 14). Taylor (1961) eloquently described these terraces as being "as straight and flat as railway embankments". In most cases the terraces are paired reflecting alteration of the erosional competence of the streams. The relationship of the terraces to other landforms such as meltwater channels and lacustrine plains indicates that the periods of downcutting were caused by successive lowering of the topographic controls produced by downwasting of the ice obstructions. The coarse texture of the compositional material (see 3.14) together with the postglacial aridity of the local climate has resulted in excellent preservation of the terrace features. The risers are steep and the tread is usually very flat. In some cases, particularly on the lower terraces, traces of an abandoned braided stream can be discerned on the tread surface. The interpretation of the terraces has been undertaken by several researchers because of the conspicuous nature of the features. Taylor (1961) suggests that the terraces that bracket Trout Creek are kame terraces. He identifies seven different levels which he feels correlate with seven different climatic phases during the deglaciation sequence (p. 185). Nasmith (1962, p. 24) also suggests that the terrace levels in the Trout Creek Valley are of ice-marginal origin, but he does not identify nearly as many terraces as Taylor, nor does he feel that climatic change is

required to produce the terrace levels.

The compositional material and landforms associated with these features leaves little question as to their origin. They are topographically beneath the kame terraces and there exists no evidence of kettles or ice-contact surfaces along the risers of these features. Moreover, as described in 2.14, a lag deposit is found in the upper horizons of the stratified gravels that make up these landforms (see plates 15 and 16). It is evident, therefore, that the terraces represent erosional surfaces carved into an outwash plain. The individual terraces represent changes in the base level of the streams as successively lower outlets were exposed during deglaciation. A close topographic similarity exists between the terraces in the Trout Creek Valley between Faulder and Mt. Conkle, and the terraces at the northern end of the Trout Creek-Penticton Diversion. The two best-defined upper terraces in the upstream portion of the valley are at elevations of 710 and 675 meters (2330 and 2215 ft) respectively. Downstream, at the junction with the diversion channel, the two uppermost well-defined terrace are at elevations of 605 and 564.5 meters (1985 and and 1852 ft) A.S.L. respectively. In addition, the lower terrace grades into the diversion channel which suggests that this probably accounts for the downcutting that produced the second terrace level further up valley. If this is the case, the third major period of downcutting would correlate with the removal of the ice blockage into Trout Creek from the Okanagan Valley and erosion took place to a base level governed by the level of Okanagan Lake.

3.29 FANS:

In order to simplify the mapping criteria, no distinction has been made between alluvial fans, which are recent landforms graded to a present-day base level, and raised fans or glaciofluvial fans which grade to a base level higher than those which presently exist. Compositionally and morphologically they are sufficiently similar to group into one category. In all cases, except for the small fans perched on the valley wall at the junction of Shingle Creek and the Marron Valley, the fans formed during deglaciation have been reworked to some extent by modern streams.

It is evident that some fans, such as the Eneas Creek fan, extended into a glacial lake and therefore the lower portion would take the form of a delta. Because such deltas are invariably associated with fans, they have not been mapped as a separate unit.

The significance of fans is that they indicate deposition into a valley that is locally free of ice and glacial meltwater. They are early postglacial to recent landforms.

The major fans mapped are located within the Okanagan Valley. Two large fan complexes form an apron that separates Lake Okanagan from Skaha Lake. From the east, Penticton Creek and Ellis Creek (see map 2) fans have extended into the valley, and Shingle Creek has contributed material from the west. The town of Penticton is situated on these fans. The morphology of these, and other fans along the south end of Lake Okanagan, suggests that there were at least two phases to their development. The upper fan surfaces grade to the high lacustrine

benches that circumscribe the southern end of the lake. This suggests that the fan deposits graded into deltas that formed in the ice-marginal or proglacial Lake Ckanagan (Lake Penticton according to Fulton and Nasmith). Subsequently, erosion has dissected the higher surface and reworked the material into the wide alluvial fans that extend into the present lake or coalesce across the valley floor at the south end of the lake.

Two fans exist within the corporate limits of Summerland. One, locally called the Lower Trout Creek Fan, extends into the main valley from the present Trout Creek Valley. The other is a complex fan deposit that underlies the central business district of Summerland. The two fans differ in that the Trout Creek Fan is similar to the fans at Penticton. It has two major defineable surfaces that grade to different levels. The upper surface grades into the kame terrace mentioned previously (3.24). The lower, and much larger surface extends more than a kilometer across the valley floor and into Lake Okanagan.

The other fan extends from the outlet of Eneas Creek (Garnet Valley) down to the upper surface of the lacustrine cliffs in the form of a broad, flat apron of well-rounded, moderately sorted and stratified sands and gravels. Unlike the fans previously mentioned, there has been no major postglacial reworking of these sediments as the discharge from Eneas Creek infiltrates the coarse textured fan material and flows toward the lake as groundwater. The significance of this fan is that it provides important evidence of the relationship between the glacio-fluvial deposits from Eneas Creek and the controversial lacustrine

deposits that occupy the sides of the main valley.

Although the morphology of the main surface of the fan indicates subaerial deposition, the distal portion of the fan became a delta which extended into the glacial lake that occupied the southern Okanagan Valley. Clear foreset beds were observed along a post-glacial gulley in East Summerland. The lacustrine sediments overlap the foreset beds with no observable interfingering of sediments. This implies that the discharge from Eneas Creek had decreased significantly by the time the silts in Glacial Lake Okanagan were deposited. In addition, it implies that the ice surface within the main valley must have retained a north-to-south gradient until the very latest stages of deglaciation. If this were not the case there would have been no cause for meltwater to cross the divide south of Peachland into the Eneas Creek Valley and discharge back into the same valley at Summerland. A study of the upper Eneas Creek (Garnet) Valley indicates that the divide separating this valley from the main valley has an elevation of 762 m (2500 ft) A.S.L. which is more than 300 m (1000 ft) above the upper surface of the lacustrine deposits.

Flint (1935, p. 114) suggests that the origin of the silts and clays that form the lacustrine benches was transported into the main valley via the tributary valleys. He postulates this because if a finger of stagnant ice is assumed to occupy the center of the valley, the amount of material contributed from up-valley sources would not be great due to the shallow gradient of the ice surface. The fact that lacustrine deposits overly the fluvial indicates that Flint's assumptions are not entirely correct.

The upper surface of the Eneas Creek Fan shows evidence of having been deposited against stagnant ice in the vicinity of the ice-scoured bedrock to the east of the outlet of Eneas (Garnet) Valley (see map 4). Between the individual bedrock outcrops, and between Wildhorse Mountain and the major outcrop, smooth-sided shallow depressions (kettles) interrupt the flat surface of the fan. These depressions indicate that stagnant ice blocks lodged around the bedrock outcrops during the deposition of the fan material.

The only major fan that is not within the main valley is one that extends from the Shingle Creek Valley into the Marron Valley. Nasmith (1962, p. 25) suggests that this fan was deposited by a stream that flowed out of Shingle Creek and drained to the northeast into Trout Creek. As mentioned in 3.27, there was possibly some flow in this direction that was routed through the Trout Creek-Penticton Diversion but little geomorphic evidence exists to show that a flow exited into Trout Creek. Instead, it appears that the fan formed as the discharge from Shingle Creek flowed across the Marron Valley to the east side of the valley where it incised a bedrock canyon that carried the flow southward along the present Shingle Creek Valley. Postglacially this canyon was abandoned in favor of a stream channel which has formed along the west side of the valley.

3.210 MODERN STREAM CHANNELS AND FLOODPLAINS:

Trout Creek is typical of the tributary streams that flow from the west into the southern end of Lake Okanagan. Within the glaciated valley area, the stream has adopted a meandering pattern in the glacial outwash. Where the stream transects the upland areas, the valley is steep-sided and bedrock controlled. The stream is contained within a tight bedrock valley where it enters the study area, but broadens to a meandering pattern from the junction with Darke Creek (Meadow Valley) to Mt. Conkle. From the north end of Mt. Conkle to the Trout Creek Fan the stream is again contained within a bedrock gorge. The stream pattern indicates that Trout Creek is underfit between Darke Creek and Mt. Conkle. This observation is supported by the existence of a wider floodplain approximately 3 m (9.8 ft) above the existing Trout Creek Floodplain. In addition, the fact that the stream has not incised to bedrock in this area indicates that the valley has probably been overdeepened by glacial erosion.

CHAPTER FOUR

4.0 INTERPRETATION:

The landform assemblages found within the study area are not similar to the assemblages described in more alpine areas of the Western Cordillera (Armstrong et al., 1965; Hughes et al., 1969; McPherson, 1970; and others). The interpretation of the landforms and glacial chronology must therefore be based on somewhat different evidence than studies carried out in areas of greater relief. In areas of high local relief moraines (terminal, lateral and recessional) are common evident landforms which can be used to trace the position of the ice front during the retreat. In the Southern Interior of British Columbia moraines are almost entirely absent (Mathews, 1944; Nasmith, 1962; and Fulton, 1967). In high relief environments meltwater plays a relatively small part in the production of landforms. In the Okanagan area meltwater is involved in the formation of seven of the ten landform units mapped. It is by reconstructing the meltwater drainage pattern that statements concerning the location of the ice margins can be made. Kame terraces and ice-marginal channels become the analog of morainic landforms in the interpretation of ice margins.

The absence of trim lines and zones of differential erosion at higher elevations suggests that none of the upland areas extended above the ice sheet during the maximum advance of the last glaciation.

It also is evident that stagnation occurred in the upland areas prior to the exposure of the topographic peaks. The magnitude of the erosion produced by this ice advance has left little, if any, evidence of earlier glaciations or the earliest phases of the last advance. The Glacial Map of Canada (1958) indicates that the ice probably reached a maximum elevation of approximately 2286 m (7500 ft) A.S.L. over the Southern Interior. This implies that the highest point within the study area was overriden by at least 943 m (3093 ft) of ice. The shape of the main valley indicates that the rate of ice movement and the thickness of ice was greater in the north end of the valley than the south (Fulton, 1972). This would partially explain the overdeepening of the valley near Armstrong. The ice-scored and erosional landforms indicate that the major ice advance was in a north-to-south direction along the Okanagan Valley with minor amounts flowing through similarly oriented smaller valleys such as the Marron Valley.

In areas of low relief the flow of ice depends on the thickness of the ice mass and it is only when the ice thins to the point that it loses its plasticity that stagnation will occur. In areas of moderate relief the flow of ice is dependent upon the thickness of ice relative to the relief (Davis and Mathews, 1944). It is therefore not surprising that the pattern of deglaciation of the study area was essentially one of ice stagnation. As soon as the upland areas were exposed, or nearly exposed as the evidence seems to suggest, the resistance to flow would have been sufficient to stop the supply of ice into the valleys so the valley lobes became

stagnant despite the fact that they contained thick masses of ice. It is important to understand however, that although the valley lobes were separated from the source region, the ice cannot be thought of as being dead ice. The lobes would have been capable of local plastic flow. Nasmith (1962, p. 9) considers that ice stagnation occurs when the surface of the ice has no gradient. This would be defined as dead ice in this study as the gradient observed on ice marginal channels and kame terraces clearly indicates that the ice surface within the study area retained a significant gradient long after the southern flow of ice had ceased.

The conclusions presented by Fulton (1967, p. 27) indicate that the flow of ice over an area of intermediate relief is largely influenced by the resistance of the underlying topography, making it logical that the lower major ice corridors such as the Okanagan Valley should continue to transport ice long after the ice in the upland areas had stagnated. Similarly, valleys that were connected to the main valley by low divides would continue to receive ice from the main valley while adjacent, but unconnected, valleys became inactive. The Darke Creek (Meadow) Valley is an excellent example of this. The divide between the northern end of Meadow Valley and the Okanagan Valley is located about 6.5 km (4 mi) south-southwest of Peachland. The elevation of the divide between the valleys is approximately 732 m (2400 ft) A.S.L. Ice-polished bedrock at the entrance to this valley (discussed in 3.21) and thin till deposits along the floor of the upper end of the valley indicates that active ice was

the last major geomorphic agent at work in the area.

The fact that this valley was supplied with ice until a later stage of deglaciation resolves a major problem in the interpretation of the study area. As mentioned, the deglaciation pattern was generally one of downwasting of ice masses and the valley bottoms were the last areas to be occupied by ice. Yet the Trout Creek Valley has been infilled with deep outwash deposits that form a train down the valley from the confluence with Meadow Valley on. Paleocurrent evidence indicates that the outwash train formed from a source up-valley but there is no indication of an area where a block of ice stagnated within the Trout Creek System. It appears that the lobe of ice that occupied the Darke Creek (Meadow) Valley was responsible for providing the meltwater and sediment that formed the Trout Creek Outwash Train.

In general, all of the valleys within the study area contain kame terraces, kettled outwash or kames. Such landform assemblages, according to Flint (1971) and Embleton and King (1968), are indicative of relatively inactive or stagnant ice masses. The exception to this pattern is found in the Three Lakes Valley where the entire valley is mantled by a thin layer of till. The elevation of this valley (nearly 792 m (2600 ft)), its north-south orientation, and unobstructed entrance into this valley from the main alley provides evidence as to why this valley alone was deglaciated while the local ice was still active. It would appear that the ice mass that occupied the main valley must have remained active until the surface of the lobe dropped

below the level of Three Lakes Valley. The ice would therefore simply thin as the surface of the main lobe downwasted and eventually the divide at the entrance to the valley would prevent the entry of ice and the valley would become ice free. This model explains not only the till within the valley, but the conspicuous lack of glacio-fluvial deposits within the valley.

Two generalizations can now be made concerning the deglaciation of the study area. First, ice stagnation was more common than ice retreat. This however, does not imply that the surface of the ice was flat and downwasting progressed equally over the entire study area. The second point is that deglaciation was rapid and continuous. The defence of this claim rests on the absence of landforms indicative of still-stands or readvances. In addition, the high rates of sedimentation evidenced by the thick rhythmites found in the lacustrine sediments (3.12) indicate a rapidly melting ice body. Although the lacustrine deposits are very extensive within the main valley, there appear to be no beach deposits or strandlines which suggests that either the levels of Glacial Lake Okanagan (Lake Penticton) during the high-stand period fluctuated considerably or the duration of the high stand was very short. The latter explanation is the more plausible as varve data presently being studied indicates that the lacustrine deposits were laid down in less than 200 years (M.J. Kent, pers. com.).

4.1 CHRONOLOGY:

The first area to be exposed during deglaciation was the upland area (1158 m - 1220 m (3800 ft - 4000 ft)) A.S.L. along the western margin of the study area. In this area thick outwash deposits with large kettles indicate that ablation of the ice sheet that covered the Southern Interior was well advanced at the time of exposure. As the ice surface lowered the exposed upland areas channelled the flow into the meltwater channel complex, part of which is presently occupied by Deschamps Creek and the upper Shingle Creek. At this stage the Marron Valley and the Okanagan Valley were still full of ice so meltwater drainage took place toward the distal portions of the ice sheet (south). During the early stages of deglaciation this would have been over the ice surface but as downwasting continued a meltwater corridor was incised into a valley that is now occupied by Brent Lake at the north end (see map 4).

The exact stage at which the ice in the upland areas became stagnant is not known but the major portion of the ice mass appears to have ceased its southward motion soon after the surface of the ice thinned to near the upper surfaces of the topographic highs. Plastic adjustments of the ice within the valleys may have taken place after this time but topographic barriers prevented the vital resupply of ice from the source areas to the north. The east-west oriented valleys were doubtlessly the first to become inactive due to their transverse orientation. The upper Trout Creek Valley west of the junction with the Darke Creek Valley illustrates this. Ice marginal

ponding took place as high as 840 m (2725 ft) A.S.L. which indicates that this valley was filled with stagnant ice at a very early stage.

As downwasting progressed, a meltwater system developed along the Marron Valley. At first the channels were formed between the ice margin and the valley walls but later the topographic valleys became the meltwater corridors. Evidence suggests that a continuous meltwater system extended from the north end of Darke Creek, through the west side of the Marron Valley and out of the study area via the Brent Lake channel. The flow through the Marron Valley appears to have passed to the west of a bedrock knob that interrupts the northern end of the valley. To the east of this obstruction a saucer-shaped depression containing lacustrine sediments indicates that a block of stagnant ice blocked the valley for some time (discussed in 3.26).

As the ice within the tributary valleys downwasted below about 730 m (2400 ft) A.S.L. the Marron Valley was abandoned as a meltwater corridor and the Trout Creek gorge along the west side of Mt. Conkle was occupied. It is not known whether or not this was a preglacial watercourse but local topography gives the impression that Trout Creek may have entered the Okanagan Valley along the Prairie Creek Valley prior to the last glacial advance. This topographic continuation of the Trout Creek Valley did not become a postglacial watercourse because of a lobe of ice that extended up the Prairie Valley and obstructed drainage through the area. The kame ridge complex (discussed in 3.23) provides evidence of this ice lobe. Trout Creek was therefore forced to incise, or re-incise, the gorge along

Mt. Conkle.

The re-routing of Trout Creek through the gorge around Mt. Conkle provided a new base level for the stream thereby causing a period of downcutting of the outwash train that infilled the valley. For some reason, probably an ice obstruction, Trout Creek does not appear to have drained southwest from Mt. Conkle into Shingle Creek via the interconnecting valley. Instead, the glaciofluvial deposits indicate a flow into the Trout Creek-Penticton Diversion. When this channel became established the discharge from the Shingle Creek system transected and reworked the outwash train that extended down the Marron Valley and flowed out through this diversion channel. Remnant terraces of the earlier deposits can be observed at the juction of the two valleys. A large fan was produced from the reworking of this material and this fan presently extends across the floor of the Marron Valley at the junction of the two valleys. After the formation of the fan it appears that the flow from Shingle Creek abandoned the Trout Creek-Penticton Diversion and incised a channel that flowed south along the existing lower Shingle Creek Valley.

The abandonment of the Trout Creek-Penticton Diversion took place when the local ice surface dropped beneath 550 m (1800 ft) A.S.L. Meltwater and the discharge from Trout Creek then began to flow along the margins of the main valley resulting in the deposition of sands and gravels in the form of kame terraces. The general elevation of these deposits is between 460 and 545 meters (1500 ft - 1790 ft) A.S.L. It is possible that it was this alteration of the Trout Creek system

that resulted in the formation of terraces as discussed in section 3.28.

Ice occupied the lower areas adjacent to the Okanagan Valley such as Prairie Creek and the small valley to the west of Giant's Head until a very late stage of deglaciation. These lobes and the blocks of ice against the bedrock exposures east of the entrance to the Eneas Creek Valley, downwasted in situ. Local meltwater ponding is evident in the Prairie Valley and Giant's Head area.

The last stage of deglaciation is the one that has generated the most confusion. The magnitude and topographic position of the silt cliffs that circumscribe the portion of Lake Okanagan that lies within the study area have generated a lot of controversy as to their origin. It is known that they are lacustrine deposits because of the texture and structure of the sediments but the environment under which they were deposited is hotly debated. Flint (1935) and Nasmith (1962) favor the hypothesis that the deposits were laid down in ice-marginal lakes as discussed in section 3.26. Although this model answers the question of the elevational difference between the lacustrine deposits and the present lake level, it is rather implausible because it requires the existence of a long, thin tongue of very slowly downwasting ice that is bracketed on both sides by deep ice-marginal lakes.

The second hypothesis is that the silt cliffs represent a higher level of a proglacial lake with the lacustrine material derived from the retreating ice front. The deposits would then have filled the entire valley to the top of the present cliffs and were

subsequently eroded out of the center of the valley by a rapid drainage of the lake. The problem with this model is there is, at present, no hard evidence to support the argument regarding a rapid flushing of the valley. The obstruction near McIntyre Bluff may have been the mechanism responsible, but further work is required to substantiate this. Additional work on the subaquatic terraces within the present lake could also lend considerable support to the argument of post-glacial flushing of the sediments.

A third possibility is that lithologic flexure caused the high stand of Lake Okanagan. As the ice melted out of the main valley proglacial ponding occured due to the differential downworping of the valley from south to north. As isostatic rebound took place the lacustrine sediments could have been elevated to their present position. The problem with this model is that an explanation must be provided as to why the lacustrine benches are confined to the south end of the present Okanagan Lake and to Skaha Lake rather than extending throughout the Valley. Moreover, it is not obvious that lacustrine cliffs would form from a gradually changing base level as would be produced by isostatic rebound.

CHAPTER FIVE

RELATIONSHIPS TO PREVIOUS STUDIES

The results of this study agree with many of the statements that have been made by previous researchers. Nasmith's work (1962) is general and because of this certain inaccuracies have been discovered. On the other hand, his study must be commended on the degree to which it goes beyond the problem that he was commissioned to solve and presents a glacial history of the entire Okanagan Valley.

At a more detailed level there are some significant differences between this study and those of previous researchers. Flint's 1935 study appears to have been hastily researched as some of the findings are refuted by the fieldwork that was done in the preparation of this study (see 3.26).

The surficial deposits map produced by Nasmith (his fig. 1) is in fact a landform map. Unlike the landform map included with this study, the mapping criteria he used is organized into a rough chronologic sequence which is divided into the following groups: glacial advance and earlier, glacial occupation, glacial retreat, late glacial, and recent. A similar classification was not adopted for this study for two reasons: First, it unifies the description and interpretation thereby making it difficult to envisage alternate explanations or interpretations; and secondly, it is somewhat arbitrary to fit landforms

into such a framework. For example, it is not clear what relationship exists between "late glacial" deposits in the tributary valleys and "glacial retreat" landforms within the main valley.

Other than differences of scale, many of the discrepancies between the landform map provided here and Nasmith's map are relatively minor. The deposits from Eneas Creek were mapped as outwash terraces by Nasmith whereas they are identified as alluvial fans according to this map. Nasmith makes a distinction between kettled outwash and outwash terraces, and between alluvial fans and raised alluvial fans whereas this map does not. The significant differences between this map and the one offered by Nasmith are:

1. Nasmith identifies a moraine ridge complex at the head of Prairie Valley. Compositionally this feature has been demonstrated to be a kame. It is therefore not necessary to postulate a tongue of ice being thrust up Prairie Valley during the late glacial period as Nasmith does (p. 26).

2. Nasmith maps the entire Three Lakes Valley as kame terraces and meltwater channels. This does not agree with the evidence provided by the fieldwork data for this study. The valley is almost entirely filled with till.

3. The valley that extends southwest from Mt. Conkle was occupied by a stream that flowed into Trout Creek according to the map prepared by Nasmith. Evidence previously presented indicates that this flow deflected south through the Trout Creek-Penticton Diversion at an early stage and south along

the east wall of the lower Shingle Creek Valley at a later date (see 3.27).

Research conducted by Hansen (1955) indicates the existence of arboreal pollen in the lowest organic sediments and even in the underlying silts of the Southern Interior which indicates that the climate was relatively mild at the time of deglaciation. Similarly Flint (1971) notes a bison skull in a glacial lake delta near Vernon which also indicates a temperate climate at the time of deglaciation. These findings suggest that the environment of the Southern Interior went quickly from a glacial to a temperate climate without an intervening stage of tundra or boreal conditions (Fulton, 1971, p. 19).

In terms of the deglaciation pattern for the entire Southern Interior, it is generally agreed by Nasmith (1962), Fulton (1967, 1975), Tipper (1971) and others that the Okanagan Valley was one of the principle exits for ice that accumulated in the Interior Plateau. The Okanagan Valley was not an unobstructed corridor, but rather the lowest exit in the rim that surrounds the trough of the Interior Plateau. Relief along the corridor provided resistance to ice penetration as is evidenced by the differential erosion that has taken place in the north-south oriented valleys as compared to the transverse valleys.

The center of ice accumulation within the Interior is somewhat controversial. Mathews (1944) and Fulton (1967) suggest that coelescent piedmont glaciers from the Coast Mountains, the Cassiar Mountains and the Rocky Mountains formed an ice sheet which infilled

the Interior to the point that the ice sheet itself became an area of accumulation. The ice in the basin then built up rapidly to form an ice dome which is thought to have flowed out through the surrounding mountains and south across the Southern Interior.

An alternate view held by Nasmith (1962) and supported by the findings of Tipper (1971) is that the ice advance took the form of coalescent piedmont glaciers that built up in the valleys of the interior until they were thick enough to drain southward through favorably oriented corridors such as the Okanagan. As the glaciation intensified the entire Southern Interior was overriden by ice but the flow was always from the mountains into the Interior and the ice dome stage was not achieved during the last advance.

In respect to the pattern of deglaciation, there is also some disagreement. Nasmith (1962), Mathews (1944) and Fulton (1967) suggest that the Okanagan Valley was deglaciated primarily as the result of the ablation of stagnant ice masses. Fulton (1967) envisages a general stagnation of ice in the entire Southern Interior. He feels that the flow of ice first ceased in the upland areas and later the lower lobes that occupied the main valleys became stagnant. Tipper (1971), on the other hand, suggests that large masses of stagnant ice were abandoned as the front of active ice retreated rapidly northward toward the source area. He argues that although there was a major pattern of stagnation, there was also an element of retreat of active ice which cannot be ignored to properly explain the landforms

of the area.

On the basis of the evidence collected for this study, it is not possible to determine whether Tipper or Fulton has the more accurate model of deglaciation. Both models are compatable with the evidence produced in this study. For example, it would be possible to fit the description of the deglaciation sequence of the Trout Creek Basin into Fulton's four phases of deglaciation (Fulton, 1967, p. 28, table 4), but evidence is also available to support Tipper's model. If the deglaciation model proposed in this thesis is essentially correct, the ice remained active from Squally Point north until after the ice had melted out of the southern end of the valley. Indeed, it is the existence of a large ice mass north from Squally Point and an obstruction to the south of the study area that is thought to have created the proglacial lake within which the thick lacustrine deposits were laid down. The critical evidence required to determine whether an ice dome existed during the last glacial advance will have to be obtained from further north in the Interior. As mentioned above, there is some evidence to support Tipper's model but, on the other hand, the ice dome model would better explain the rapid and uninterrupted retreat of ice from the Southern Interior.

It can be stated however, that the models proposed by Flint (1935) and Nasmith (1962) are questionable in so far as they postulate a long, flat-topped tongue of ice in their explanations of the origin of the silt cliffs that are presently found along the sides of the southern Okanagan Valley. It would seem much more likely that these

deposits originated in proglacial rather than ice-marginal lakes. However additional detailed research on the lacustrine sediments is required before the origin of the silts is finally resolved.

Plate 1

Plate 1. Kame complex at the upper end of Prairie Creek Valley. Photo taken looking south with Trout Creek gorge in background.

Plate 2. Internal composition of a section along west side of kame complex.

Plate 3. Internal composition of an east-facing section at the north end of the kame complex.

Plate 4. Internal composition of the east side of the kame complex.

Plate 2

Plate 3

Plate 4

Plate 5. Kettles in the northern (up-valley) end of a major outwash train.

Plate 6. Terraces on the northern (up-valley) end of the Trout Creek-Penticton diversion (Source: Nasmith (1962)).

Plate 7. Outwash plain along western edge of Marron Valley.

Plate 8. View looking northwest up the Trout Creek Valley from Mt. Conkle. Note kame ridge complex just to right of center and the terraces along the south side of the valley.

Plate 9

Plates 9 & 10. Silt cliffs along the margins of the southern end of Lake Okanagan

Plate 10

Plate 11. Meltwater channels which appear as notches incised into Squally Point (Source: Nasmith 1962).

Plate 12. Meltwater channel incised into the northwest facing side of Mt. Conkle.

Plate 13. View looking south across Trout Creek Valley. Trout Creek canyon is the valley to the left and Marron Valley is divided by a bedrock knoll in the center of the photo. Note the two clearly defined terraces along the southern side of the valley.

Plate 14. View looking north across Trout Creek Valley. The lower end of Three Lakes Valley is in the center of the photo and Meadow Valley is on the left side. Note two terrace levels on the north side.

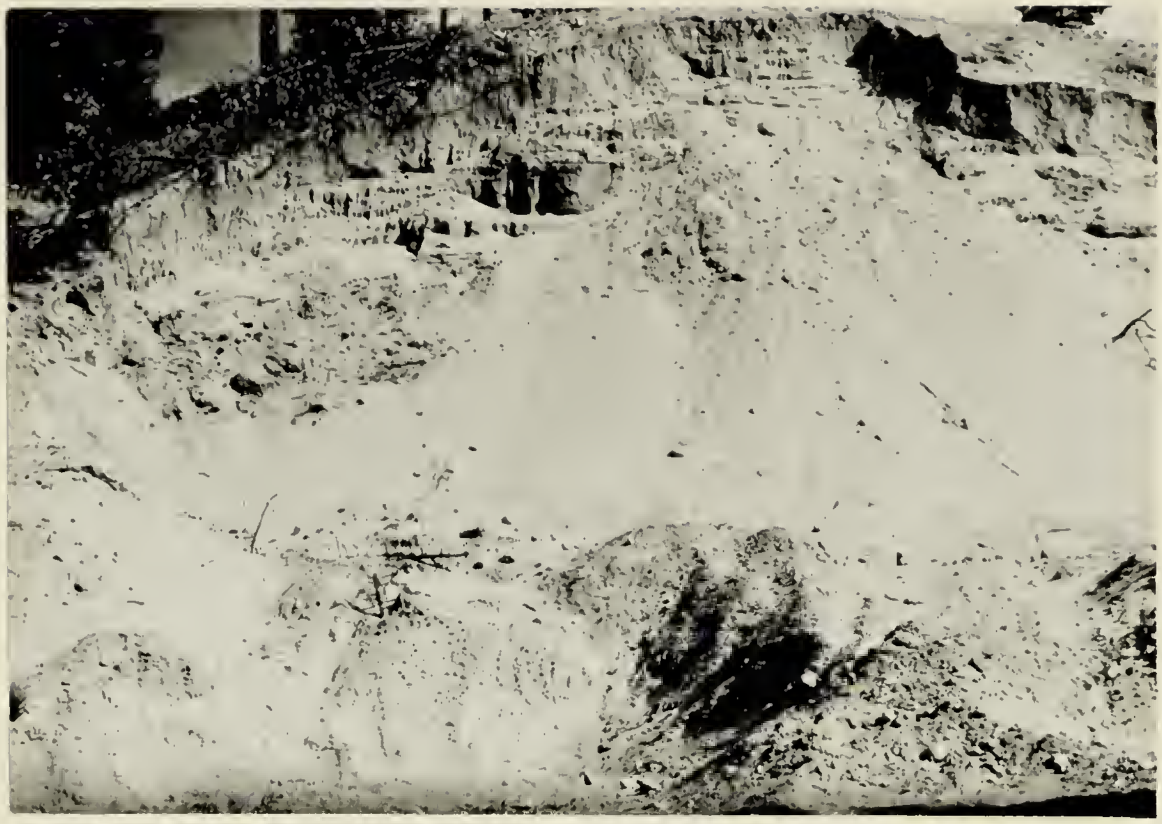


Plate 15. Exposure in upper (710 m; 2330 ft) terrace on south side of Trout Creek Valley. Note coarse texture of upper strata.

Plate 16. Exposure in lower (675 m; 2215 ft) terrace on south side of Trout Creek Valley. Note coarse texture of upper strata.

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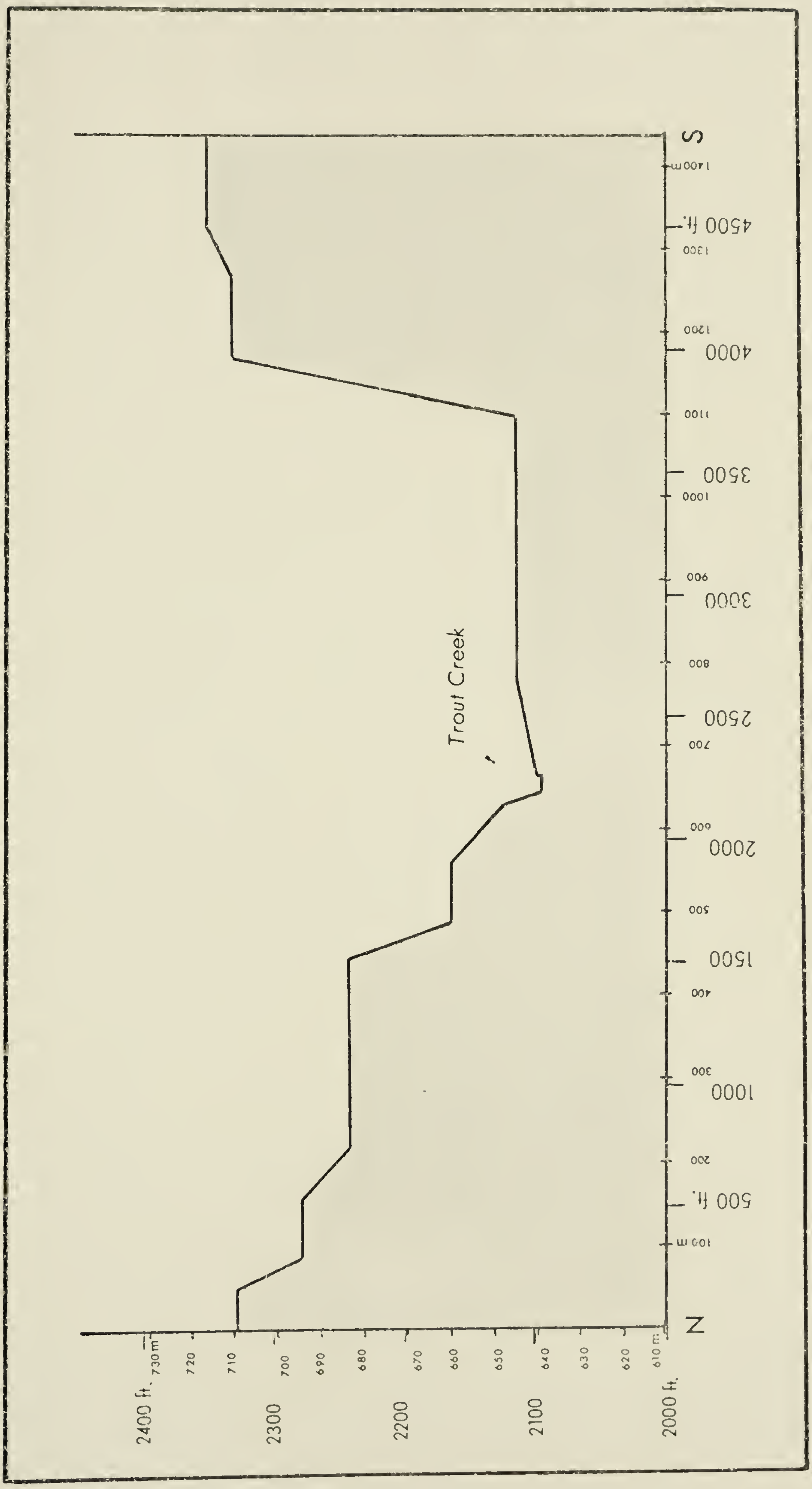
Appendix I

Cross-sectional Profile of Trout Creek Valley

Southeast of Faulder



N
S
Trout Creek
2400 ft.
2300
2200
2100
2000 ft.
730 m
720
710
700
690
680
670
660
650
640
630
620
610 m
500 ft.
1000
1500
2000
2500
3000
3500
4000
4500 ft.
100 m
200
300
400
500
600
700
800
900
1000
1100
1200
1300
1400 m